Volume 28 SEPTEMBER, 1958 Number 3

Journal of Sedimentary Petrology

A Publication of the Society of Economic Paleontologists and Mineralogists
a Division of
The American Association of Petroleum Geologists

AUG 14 1958

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FIGS. 1–4, SEPTEMBER, 1958

FENESTRATE BRYOZOAN CORE FACIES, MISSISSIPPIAN BIOHERMS, SOUTHWESTERN UNITED STATES[1]

LLOYD C. PRAY
The Ohio Oil Company, Littleton, Colorado

ABSTRACT

Preliminary data indicate that fenestrate bryozoa form the major faunal component in the core facies of the "crinoidal bioherms" of Osagian (Mississippian) age exposed in southern New Mexico. The core facies of these bioherms consists largely of aphanitic calcite and sparry calcite within a meshwork of intact and partially comminuted fenestrate bryozoans. Some of the bryozoans are in growth positions. Whether in growth position or as broken debris, the bryozoa are considered to be of significance in the genesis of these abrupt mound and ridge-shaped structures. The aphanitic calcite is interpreted as primary lime mud that was indigenous to the core facies. Much of this may be of algal origin. Algal frame-builders are not recognized. The core facies grades laterally with abruptness into a flank facies of coarse crinoidal debris, and thence into cherty calcilutites of the "interreef" facies. The core facies of many bioherms formed above the level of adjacent contemporaneous sedimentation, and some bioherms had several hundred feet of depositional relief. Most crinoidal material is considered to be indigenous to the flank or "interreef" facies. The core facies biotic assemblage is tentatively interpreted to not constitute a true reef assemblage, although apparently capable of growth into somewhat turbulent waters.

Fenestrate bryozoans are probably of genetic significance in the Mississippian "crinoidal bioherms or reefs" in the subsurface of north-central Texas, and may be important in those of northeast Oklahoma. The New Mexico bioherms appear similar to the "reefs" of the Waulsortien facies of western Europe and to some of the English reef knolls, all of comparable Mississippian age. Good preservation of the primary faunal constituents, the structural simplicity, and the excellent exposures indicate that much needed data on paleoecological interpretation of Mississippian "reef" problems should result from detailed analysis of the bioherms of southern New Mexico.

INTRODUCTION

Abrupt mound and ridge-shaped carbonate masses of Mississippian (Osagian) age that are up to 350 feet thick crop out in southern New Mexico where they were first reported and described in considerable detail by Laudon and Bowsher (1941). These structures and apparently similar ones in the subsurface of north-central Texas have commonly been referred to as "crinoidal bioherms." Although the New Mexico occurrences have been known and examined by many geologists, including the writer, since 1941, their genesis has remained obscure, particularly the nature and origin of the core facies. The most enigmatic aspect has been the identity of the frame-builders, if indeed frame-builders were ever present. The purpose of this preliminary article is to call attention to the presence of fenestrate bryozoa as the major faunal component of the core facies and to suggest their significance in forming a mesh-like framework that locally permitted growth of the core facies to heights several hundred feet above the level of the adjacent contemporaneous sedimentation. To the writer's knowledge, fenestrate bryozoa have not been previously reported in the core facies of the New Mexico bioherms or considered to be of genetic significance in any of the Mississippian "crinoidal bioherms" of the southwest.

Late in 1957 the writer examined polished surfaces prepared on two large hand specimens from a Mississippian biohermal complex of the Sacramento Mountains of southern New Mexico. An abundance of fenestrate bryozoa was noted in one specimen which on cursory inspection had appeared to be a rather unfossiliferous calcilutite. The bryozoans were in various degrees of fragmentation and in diverse orientations. They appeared to form a meshwork through the rock. Subsequently, Alan Horowitz and the writer spent five days in the field in the Sacramento Mountains in reconnaissance inspection and sampling of Mississippian bioherms to further investigate the potential significance of fenestrate bryozoans. This

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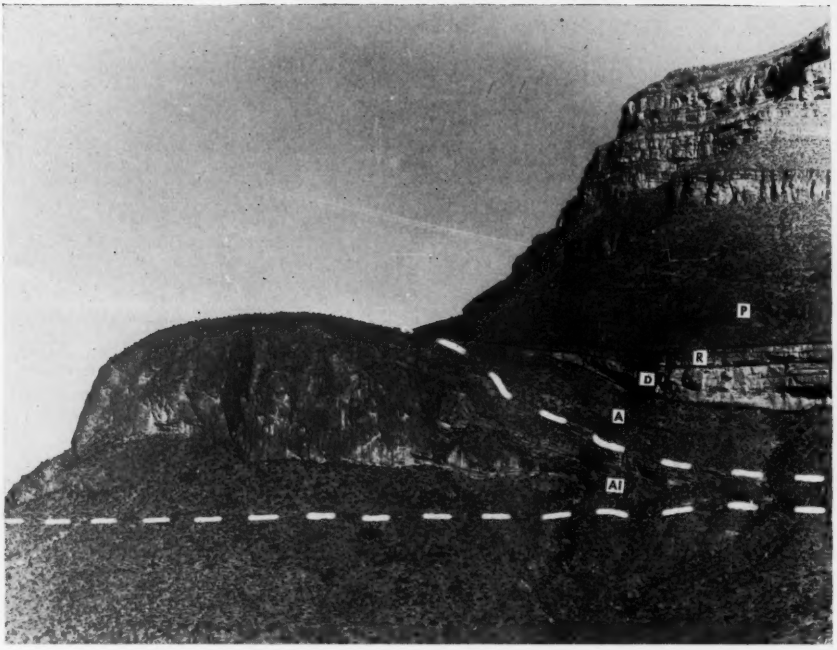

FIG. 1.—The Muleshoe Canyon bioherm, Sacramento Mountains, New Mexico. The bioherma facies (Al) is about 350 feet thick and largely composed of core facies at the center of this domica bioherm (left center of the photograph). Traced laterally, as shown by the dashed white lines, the core facies grades abruptly into coarse crinoidal strata of the flank facies, which thins to about 65 feet on the right edge of the photograph, a distance of 1000 feet from the center of the bioherm. The biohermal facies (Al) here include the Alamogordo, Nunn, and Tierra Blanca Members of the Lake Valley Formation. The Arcente (A) and Dona Ana (D) Members of the Lake Valley Formation, the Rancheria Formation (R) of Meramecian age, and the lowest of the Pennsylvanian strata (P) pinch out near the top of the resistant biohermal knob.

article is based largely on this field work and subsequent petrographic investigations. To date, six of the Mississippian bioherms of the Sacramento Mountains have been studied. These range in size from one about 25 feet thick to the largest in the area which is about 350 feet thick (fig. 1) and represent several different stratigraphic levels within the Mississippian host formation. All six show a consistent concentration of fenestrate bryozoans in the core facies and the concentration of the crinoidal limestones to the surrounding or flank facies. In addition, the only well core the writer has been able to obtain from a Mississippian "reef" or bioherm of north-central Texas is nearly identical with core facies of some of the New Mexico bioherms. Although many outcrop and subsurface examples need to be critically examined to determine the extent and significance of this type of carbonate deposit, the present evidence seems sufficient to justify this preliminary article. Detailed studies of several of the New Mexico bioherms and reconnaissance examination of some other "crinoidal bioherms" in adjacent states will be reported when the additional data has been assembled and interpreted.

In this article the abrupt depositional carbonate structures are generally referred to as bioherms (or biohermal complexes) in the sense of the definition of bioherms by Cumings (1932) and following the usage of Laudon and Bowsher (1941). Interpretive

aspects of wave-resistance or hydrographic expression are discussed briefly in the article, but use of the term bioherm in Cumings' sense (1932) carries no implication as to these interpretive aspects.

NEW MEXICO OCCURRENCES

The stratigraphic studies of Laudon and Bowsher provide most of the published data on the Mississippian stratigraphy of the part of southern New Mexico in which the bioherms are known. The bioherms are developed in the Lake Valley Formation, considered early Osagian on the basis of faunal studies. Laudon and Bowsher (1941) named three members of this formation, and in 1949 established a six-fold subdivision, summarized in table 1.

TABLE 1.—*Lake Valley Formation*

Member	"Normal" lithology
Dona Ana Member	Crinoidal calcarenite and calcirudite
Arcente Member	Argillaceous calcilutite and shale
Tierra Blanca Member	Crinoidal calcarenite and calcirudite
Nunn Member	Crinoidal "marl"
Alamogordo Member	Cherty calcilutite
Andrecito Member	Silty calcarenite

The Lake Valley Formation is commonly 200 to 400 feet thick where all members are present. Contacts between members are gradational with the possible exception of an unconformity at the base of the Arcente Member. The Lake Valley Formation overlies an unconformity separating it from the Caballero Formation of Kinderhookian age. It is overlain by Meramecian strata from which it is separated by an unconformity with an angular discordance of a few degrees. The stratigraphic sequence of major concern for the interpretation of the bioherms consists of the Alamogordo, Nunn, and Tierra Blanca Members since most biohermal growth occurs within these units. For the purpose of this article, a three-fold classification of lateral lithologic facies within this sequence is more useful than the vertically differentiated members. These are the core and flank facies, both considered parts of the biohermal complex, and the "interreef" or "normal" facies. The lithology indicated in table 1 indicates the "normal" facies deposited beyond the area of major biohermal influence. These three lateral facies were developed contemporaneously. Most of the bioherms began growth during the deposition of the Alamogordo Member. Many of the smaller structures are entirely enclosed within the "normal" facies of the Alamogordo Member, but some of the larger structures continued their upward growth during the time of deposition of the Nunn and the Tierra Blanca Members. Biohermal growth appears to have ceased prior to the deposition of the Arcente Member.

In southern New Mexico the bioherms are most abundant in exposures along the western escarpment of the Sacramento Mountains in the south-central part of the state. Here the Lake Valley Formation crops out in a north-south belt about 18 miles long. The bioherms are most numerous in the northern part of this outcrop belt directly east of the city of Alamogordo. Bowsher (1948) indicated the position of biohermal outcrops on a small geologic map. Some further geologic details of the area, a small geologic map, and road logs were published by the writer (Pray, 1954). Several of the largest bioherms occur along the mountain front about 7 miles south of Alamogordo. Two of these are large enough to be conspicuous from the main highway, 4 miles west of the mountain front. One of these is shown by figure 1.

Many of the larger bioherms of the northern part of the Sacramento Mountain outcrop belt are elongate and trend slightly west of north. Bioherms farther south appear to be domical or mound-shaped. In contrast to the sharply convex upper surface of the bioherms, the basal surface is commonly flat, or slightly convex downward. The flank strata of the bioherms dip sharply away from the upper parts of the biohermal core, commonly at angles of 30 to 35 degrees. Locally the core facies appears to have settled into the soft substrata of the Andrecito Member to a greater degree than the adjacent flank facies.

Core Facies

The core facies forms the central mass of the bioherms or biohermal complexes. The general impression of the lithology is that of a massive, dark-colored, relatively unfos-

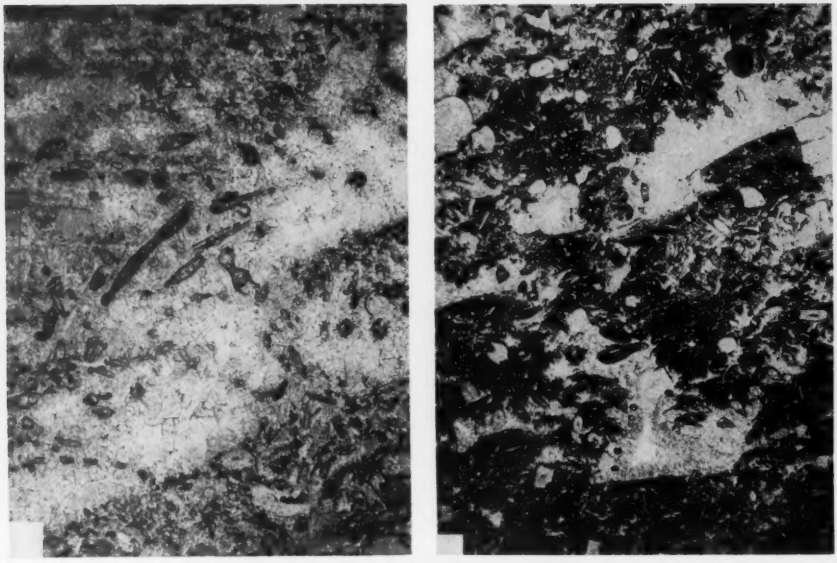

FIG. 2.—Photomicrographs of variants of core facies lithology. Nearly all of the darker masses in the left illustration are sections of fenestrate bryozoans. In the illustration on the right, some of the fenestrate material is dark and conspicuous, but much inconspicuous fenestrate debris occurs in the fine-grained darker areas. The illustrations are oriented in normal stratigraphic position, and the notch in the lower left corner is 1 mm square.

siliferous calcilutite. It has been described as black, hard, almost featureless limestone . . . distinctly and markedly unfossiliferous. Although darker than most of the enclosing crinoidal limestones, fresh specimens of the core range in color from medium gray to light olive gray with lighter shades dominant. Critical inspection, especially of thin sections or polished slabs, reveals that fenestrate bryozoans form the most abundant faunal component. In most specimens they are more abundant than all other faunal elements combined. Although fenestrate bryozoans rarely form more than 20 percent of the rock, this amount appears sufficient to be of major genetic significance. Some of the fenestrate fronds are relatively intact and in growth positions; others have been broken and deposited as fragmental material. Whether erect and thus forming a current barrier and effective sediment trap or deposited as broken mesh-like zoarial fragments and forming a sediment-retaining mat on a depositional surface, this material provides a framework for the build-up of the core facies. Fenestrate bryozoan debris that consists of only a few cells or of even smaller pieces is common. This material furnished detritus for the biohermal complex, but debris smaller than the zoarial mesh size would have little or no sediment-retaining or frame-building ability. Figure 2 shows two photomicrographs of typical core lithology. The excellence of preservation of primary features of the fenestrate bryozoans is indicated by the two photomicrographs of figure 3.

Echinodermal segments, presumably largely crinoidal columnals, are the most obvious faunal component of the core facies in the field, because of their ease of recognition. They are less abundant than bryozoa and almost entirely lacking in some specimens. Much of this material occurs as individual columnals or short sections of columnals, scattered like "plums in a pudding." Thin lenses and pockets of sorted crinoidal debris occur locally. These increase toward the flanks, as noted by Laudon and Bowsher (1941, p. 2129). Many of

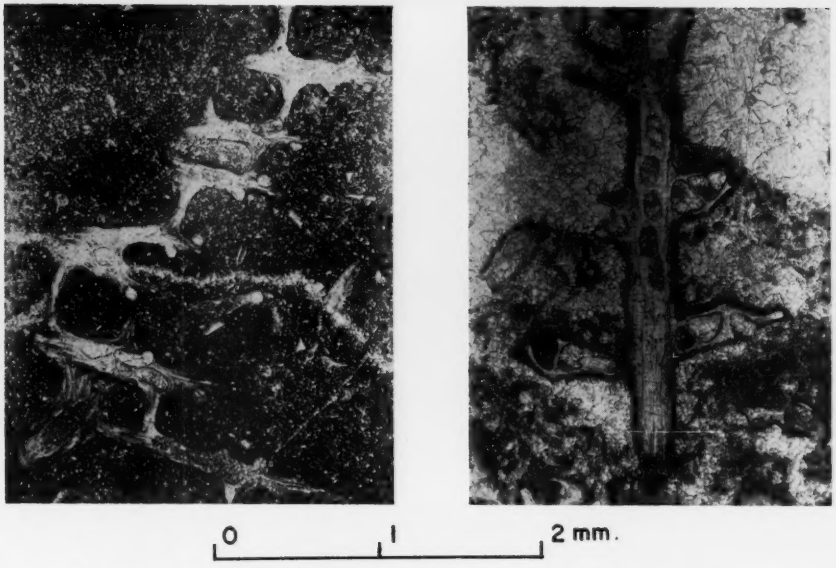

FIG. 3.—Photomicrographs of fenestrate bryozoan material in bioherm core facies. Note the preservation of structural details in the bryozoans.

these may represent temporary incursions of the flank facies but others probably reflect material indigenous to the core facies.

Recognizable faunal components other than fenestrate bryozoans and echinoderms are quantitatively insignificant, forming less than one percent in some rocks. Other types of bryozoans, brachiopods, orthoconic cephalopods, and small corals are locally observable in the field. Trilobite fragments and ostracods are commonly noted in thin sections.

Aphanitic calcite and more coarsely crystalline (sparry) calcite form two-thirds or more of the volume of the core facies as shown by figure 2. On the basis of preliminary study, the aphanitic calcite (*ca.*0.01 mm) is interpreted as derived from primary lime mud, and the sparry calcite is considered to have formed by infilling of primary voids in the rock and by local recrystallization of lime mud or other primary constituents. Aphanitic calcite predominates in most specimens that have been examined. The aphanitic and sparry calcite appear to vary inversely in abundance. Where extensive recrystallization has not occurred, the amount of aphanitic calcite may indicate the turbulence of the depositional environment; the lower the content, the more turbulent the environment.

The origin of the lime mud is obscure, a situation common to most ancient and Recent carbonate deposits. Some undoubtedly represents comminuted faunal or floral constituents broken by either mechanical or organic processes. Inorganic precipitation is a possible source that cannot be evaluated from the present data, although it seems unlikely to have been of major significance. The writer believes that much of the lime mud was of algal origin, either as carbonate needles secreted in an organic framework (Lowenstam, 1955) or as an algal slime formed on or near algal plants and precipitated as the plants extracted carbon dioxide from the immediately adjacent sea water. Distinct algal structures have not yet been recognized in the core facies unless one accepts the interpretation of Elias and Condra (1957, p. 43) that part of the fenestrate bryozoan lattice is algal in origin. The writer formerly believed that the lime mud of the core facies must have served a cohesive

function to permit formation of the abrupt mounds, possibly through possessing a filamentous nature prior to recrystallization. Perhaps this is true and the evidence has not yet been observed, but the abundant presence of a fenestrate bryozoan meshwork renders such an explanation less necessary.

The site of origin of the lime mud seems more certain than its mode of origin but is also somewhat obscure on the basis of present evidence. The writer believes that much of it is indigenous to the core facies of the bioherms rather than having been formed elsewhere and transported to the site of the bioherm. Reasons for this are most applicable to the larger bioherms that rose well above the level of the adjacent sedimentation and in which there is a much greater amount of lime mud than in the adjacent contemporaneous facies. If the lime mud were not indigenous to the core facies, it would require transport and selective deposition on the biohermal hydrographic prominence. Several mechanisms could be considered. The lime mud could have been deposited by settling, as a "lutite veil," from suspension in surficial or intermediate depth waters; however, an equal amount or more would logically be expected in the adjacent lower areas. If the transport were by bottom-hugging turbidity currents, deposition in the low areas would again exceed that of the biohermal high. Another transport mechanism would be by suspension or traction by bottom currents other than the episodic turbidity type, but again it is difficult to get the lime mud deposited selectively on the biohermal high. A fourth mechanism, more plausible than the preceding three, would be to have the lime mud carried in suspension in waters that circulated past the biohermal sites and have some of the mud selectively removed at the bioherm. This could happen by a filtering or current-checking function that fenestrate bryozoans could provide or possibly by the trapping of lime mud in surficial ooze of algal origin as proposed by some for the growth of algal stromatolites. However, distinct layered structures common to algal stromatolites have not been recognized. It seems more reasonable to reject mechanisms requiring transportation of lime mud to the site and to consider instead that it formed essentially in place.

Flank and "Interreef" Facies

The flank facies are those occurring between the core facies and the "interreef" or regionally "normal" facies. Data regarding the flank and "interreef" facies have been given in considerable detail in the cited articles of Laudon and Bowsher (1941, 1949); details are beyond the scope of this article. Laudon and Bowsher's studies were particularly detailed in the macro-paleontology of those units where recovery of relatively complete fossils such as crinoid calices is possible, as with the crinoidal "marls" of the Nunn. The Nunn Member, which represents both flank and "interreef" facies, provided the major faunal collections.

In sharp contrast to the core facies, the flank facies is composed almost entirely of bedded calcirudites and calcarenites of echinodermal material. These materials are commonly cemented with sparry calcite. The present data suggest that most of the echinodermal debris that forms the flank lithology was indigenous to the flank environment. The outer flanks, however, may have formed largely from debris transported from growth positions closer to the core edge of the flank facies. Fenestrate bryozoa, largely debris, occur interstitial to coarser echinodermal debris and is especially common near the core facies. Coarse limestone conglomerates, with clasts resembling the core facies and a matrix of echinodermal debris, form conspicuous local units. These probably are slump deposits, perhaps triggered by unusually deep wave action. Aphanitic calcite, interpreted as primary lime mud, is not common in the flank facies, but where present appears more characteristic of the outer portion than of the part near the core facies. The same is true of argillaceous material.

The "interreef" facies represents the strata deposited contemporaneously with the core facies but in areas beyond the influence of the growing bioherms. Where the bioherms are developed entirely within the Alamogordo Member, the "interreef" facies consists of the characteristic medium gray, cherty calcilutite, generally about 40 feet thick in the Sacramento Mountain area. The strata are resistant to weathering and commonly form a low scarp. The bedding planes are sharply defined, though somewhat un-

dulatory, and generally ½ to 2 feet apart. The rock consists largely of aphanitic calcite and contains a sparse fauna. Occasional intact fronds of fenestrate bryozoans and other evidence indicate a very quiet-water environment. A few small bioherms occur entirely within the crinoidal calcirudites and calcarenites of the Tierra Blanca Member; in these situations, the flank facies and the "interreef" facies are lithologically similar. Where biohermal growth continued from Alamagordo time into the time of deposition of the Nunn and Tierra Blanca Members, the "interreef" facies consists of cherty calcilutites overlain by crinoidal "marls," calcarenites, and calcirudites.

Hydrographic Expression

Many of the bioherms are interpreted to have formed abrupt prominences on the sea floor at the time of deposition. This interpretation was first made by Laudon and Bowsher (1941, p. 2129) and is confirmed by subsequent work. Evidence for this is found in the gross stratigraphic relationships of the core facies to the flank facies, particularly in the thickness contrast between these facies. The best single example is afforded by one of the largest bioherms in the Sacramento Mountains, which crops out in the center of sec. 28, T. 17 S., R. 10 E. on the north side of a small canyon (locally referred to as Mule-shoe Canyon). This exposure was illustrated by Laudon and Bowsher (1941 p. 2149). Figure 1 shows the topographic prominence or klint formed by the resistant core facies of this bioherm as viewed from the southeast. The massive core facies grades abruptly toward the east (right on the photograph) into distinctly bedded flank facies. The biohermal core is essentially domical in shape and the writer considers the present top of the klint to be near the top of the original bioherm. Thus the prominent cliff face of figure 1 is a cross-section several hundred feet south of the center of the core. From the base of the Alamogordo Member to the present top of the klint is approximately 350 feet. This vertical distance affords a minimum thickness for the core facies. Although not readily discernible on the illustration, core facies crops out below the massive cliff on the western half of the klint. Small biohermal masses and the abundant crinoidal limestone (flank facies) that occur in the lowest part of the Alamogordo Member in the area of the photograph are evidence that the core facies persists downward to the base of the Alamogordo Member in this immediate vicinity, probably under the center of present klint. The original thickness of the consolidated core facies may have been thicker than the present 350 feet, since the present top of the klint extends above the level of two unconformities, one separating the Osagian and Meramecian formations and the other the Mississippian from the Pennsylvanian.

The strata intertonguing with the core facies can be traced laterally toward the much thinner flank facies along the slope shown in figure 1. The exposures in the gully on the right side of figure 1 are about 1000 feet from the inferred center of the bioherm. From the gully, and continuing to the east (right on the photograph) for a quarter of a mile, the total thickness of the entire stratigraphic interval between the base of the Alamogordo Member and the base of the Arcente Member is about 65 feet. This is less than one-fifth the thickness of the adjacent core facies. An additional factor that makes the thickness contrast even greater is that the uppermost part of the 65-foot thick flank section of this bioherm appears to be traceable into strata of the upper part of the Tierra Blanca Member, that is, into strata that are younger than the present core facies. To consider the possibility of compaction only makes the primary thickness contrast greater, since the sorted crinoidal calcarenites and calcirudites that form the bulk of the flank facies have not undergone much compaction. If compaction has appreciably affected the thickness contrast, it is probable that the core facies, with a high content of primary lime mud, compacted more than the flank facies.

The marked thickness contrast between the core facies and the flank facies necessitates the interpretation that the core facies of this bioherm stood well above the level of adjacent contemporaneous sedimentation. It seems probable that the topographic relief at the end of the growth of this bioherm was as much as 250 feet in a lateral distance of 1000 feet. Similar thickness contrasts and the high dips present in the non-compacting flank facies adjacent to the cores of other bioherms suggest that the core facies gen-

erally formed above the level of the flank facies. For smaller bioherms, the differential elevation was correspondingly less, and for some bioherms may have been only a few feet.

Demonstration of the hydrographic relief of this bioherm (fig. 1) permits many interesting paleoecologic interpretations, particularly those pertaining to the depths at which the flank and "interreef" facies were deposited. The flanking crinoidal facies formed in deeper water than the core facies, and the outer part of the flank facies was probably deposited at depths of the order of 300 feet or more. These bedded crinoidal limestones have many similarities to crinoidal limestones of areas where many geologists have inferred a much shallower depth. The indigenous faunas of the flank and "interreef" facies at critically selected localities can be studied with reasonable assurance of a minimum water depth at the time of formation equal to the height of the biohermal core. This can be applied most rigorously to the upper beds of the sequence that can be traced into the uppermost part of the core facies, but it may prove applicable to the entire thickness of the outer flank and interreef facies. As the bioherms are not believed to have grown into the zone of vigorous wave action, the minimum depth figure can be increased by several tens of feet (Dietz and Menard, 1951).

MISSISSIPPIAN BIOHERMS OF OTHER AREAS

Mississippian bioherms of Osagian age that others have considered similar to those in southern New Mexico occur in the subsurface of north-central Texas and in outcrops of northeastern Oklahoma. In view of the long delay between knowledge of the existence of the bioherms and recognition of a bryozoan core facies in the New Mexico area, it is interesting to speculate on the possibility that these other occurrences are of similar type.

Bioherms presumed to be similar to those of New Mexico occur in the sub-surface of north-central Texas on the eastern side of the Bend Arch in the Chappel Limestone of Osagian age. The regional distribution and some details have recently been summarized by Turner (1957, p. 61, 62) as follows:

"Through an arcuate belt extending from Comanche County on the south to Montague County on the north, the Chappel seems to form the platform from which grew the crinoidal bioherms of the Mississippian reef terrain. These scattered reef masses occasionally developed to a height of about 500 feet. Their flanks have relatively steep slopes and their areal extent is usually less than one square mile. Studies of local reef masses suggest that they may be somewhat similar in morphology to the Mississippian Lake Valley bioherms exposed in the Sacramento Mountains of New Mexico."

Well cores from these sub-surface structures appear to be scarce. To date, the writer has obtained only a single piece, six inches in length. This was cored from near the center of a 300-foot thick section of Chappel Limestone in the approximate middle of the bioherm that forms the La Pan Mississippian field of Clay County, Texas, near the northern end of the belt discussed by Turner. The core specimen (fig. 4) consists largely of fenestrate bryozoan mesh-work, with minor crinoidal columnals and other fauna in a matrix of sparry calcite. The bryozoans are in diverse orientations, and many fronds are relatively intact and considered to be in growth positions. Some appear somewhat cup-shaped, and appear to have trapped debris within the upright cup. The similarity of the texture and biotic constituents of this specimen to the core lithology of the New Mexico bioherms of the same size is striking. The major difference lies in a greater proportion of sparry calcite to aphanitic calcite in the Texas specimen compared to the average New Mexico bioherm, but these same proportions are locally observable in the latter where depositional conditions are interpreted to have been particularly turbulent. Much of the sparry calcite of the core specimen from Texas appears to have grown inward by open space crystallization that began at a bryozoan frond, interpreted to have formed the wall of an original cavity of the rock. Similar calcite in reef rocks was reported by Newell (1955, p. 308). The resultant texture also resembles some of the "tufa" bands that are such a moot point in interpretation of the lithology of the English reef knolls. Although one piece of core is admittedly very meagre evidence to characterize the

FIG. 4.—Vertical surface of well core taken from near the middle of a bioherm forming the La Pan Mississippian field, Clay County, Texas. Fenestrate bryozoans form nearly all of the dark spots, rods, and mesh-like structures, and form much of the lighter gray debris in the lower right quarter of the illustration. Note the diverse orientation of many relatively intact bryozoan fronds, and the apparent influence of these fronds on the crystallization of the sparry calcite. Core specimen from 6289–6320 foot depth in the Panhandle and Lagorce No. 1 Lyles well.

north-central Texas bioherm structures, it suggests the possibility that further studies will show fenestrate bryozoan core facies to be a feature of most of these bioherms.

Crinoidal bioherms exposed in northeastern Oklahoma in the St. Joe Member of the Boone Formation (Osagian) were reported by Laudon (1939) and have recently been studied by Harbaugh (1957). Both report that these bioherms consist largely of crinoidal debris in both the rock of the more massive cores and in the bedded flank deposits. The cores are 10 to 40 feet in thickness and from 50 to 1000 feet in width (Harbaugh, 1957, p. 2544). Fenestrate bryozoans are not cited as faunal constituents of these bioherms by Laudon or Harbaugh, but in a letter to the writer of January 31, 1958, Harbaugh states:

"In northeast Oklahoma, the massive bioherm cores do contain much crinoidal material. Other material is also present, such as fragments of fenestellate bryozoans. The grain size in the limestone in massive cores varies—from very fine to moderately coarse, and the most obviously crinoidal material is found in the thin-bedded limestones, which might be interpreted to be flanking deposits."

The writer has not visited the Oklahoma localities, but Harbaugh kindly furnished two specimens collected from the massive cores of two bioherms. One is essentially a clean, crinoidal calcirudite. The other contains abundant fenestrate bryozoan debris and aphanitic calcite, together with crinoidal and other fossil debris. It seems possible that a core facies similar to the New Mexico bioherms occurs in those of northeast Oklahoma. If so, either the vagaries of dissection has not exposed them or the core lithology has not been examined critically, a situation of which the writer was guilty for a period of 10 years with respect to those of southern New Mexico. If the core facies were relatively small compared to the flank facies, perhaps most exposures would not intersect the true core, and thus the visible cross-sections of the mounds would consist largely of the flanking crinoidal debris. In this connection it is perhaps significant that in some of the New Mexico bioherms, such as shown in figure 1, the change from distinctly bedded to massive limestone occurs within the crinoidal limestones of the innermost flank facies, rather than at the nearby transition from crinoidal limestones to a typical core facies lithology. The existence of steep dips in the crinoidal strata at distances of 400 feet (Harbaugh, 1957, p. 2534) from the visible massive core is suggestive evidence that a core facies, possibly different in lithology from that now considered as core facies, existed at a topographically higher level than is now exposed.

The Borden Group bioherms of Indiana (Stockdale, 1931) have been well studied and found to consist largely of crinoidal material. Fenestrate bryozoans are common minor faunal components. Although these bioherms are about the same age as those of Oklahoma, Texas, and New Mexico, they apparently are of a different type. The Borden bioherms are interpreted to have not formed depositional mounds on the sea floor but to have been simply localized crinoidal meadows on a sea floor that was essentially level. Perhaps a transitional series exists between the Borden type and those that formed abrupt, high, depositional mounds typified by the larger bioherms of New Mexico. In this series, the Osagian bioherms of Oklahoma may prove to be intermediate types. Additional comparative studies are needed to evaluate this speculation.

BIOHERM OR REEF?

The New Mexico structures have been termed bioherms by most geologists who have examined them. Those in north-central Texas have been variously referred to as bioherms or reefs. If "bioherm" is used in the sense discussed by Cumings (1932) there is little doubt as to its applicability. Those of the Lake Valley Formation in New Mexico and apparently similar ones of adjacent areas are carbonate mounds or lenses which were formed largely by sedentary organisms and that differ in lithology from the enclosing strata. Cumings (1932) applied the term bioherm to a wide variety of genetic types of carbonate masses. He specifically included coral and algal reefs, such as those of the Pacific Ocean area, algal banks, and the carbonate accumulations which are interpreted to have not formed prominences above the sea floor at the time of deposition, such as the *Lucina* or tepee buttes of Colorado and the crinoidal mounds

of the Borden Group of Indiana. Moreover, the genetic interpretation as to whether the carbonate mass was formed by a biotic assemblage of wave-resistant potential was neither stated nor implied in Cumings' (1932) discussion. Although various authors have since attempted to restrict the usage of "bioherm" to a narrower sense, the main utility of the term appears to be its very few genetic restrictions. Often the genetic aspects can only be determined by thorough study of specific structures, and a "wastebasket" term serves a useful function. Thus, the use of bioherm for the Mississippian carbonate masses of the type exposed in New Mexico appears justified, but it does not greatly clarify their genesis. Their genetic type is better understood if they are placed in a category of bioherms interpreted to have formed prominences that rose above the general level of the sea floor, a type for which no specialized term in common usage is available.

Whether the New Mexico bioherms can be called true reefs in the sense discussed in detail by Lowenstam (1950) is a significant question. This definition of a reef requires a biotic assemblage that has the potential of forming a wave-resistant structure and of being capable of net growth within the zone of violent wave agitation. Growth into a zone that is only occasionally subject to wave-induced turbulence would fall short of strict adherence to this concept of a true reef assemblage.

The larger of the bioherms in the Sacramento Mountains and probably those of north-central Texas appear to have grown upward into a somewhat turbulent environment. This is indicated by lithologic units in the core facies containing well-sorted coarse debris and almost lacking in fines which apparently were winnowed out by current action. The presence in the flank facies of limestone conglomerates containing coarse clasts interpreted as core lithology is evidence for some erosion of the core facies, possibly during times of unusual wave turbulence although these may be of slump origin triggered by processes unrelated to wave-induced turbulence.

If algae were important organic constituents, it seems probable that the core facies biotic assemblage grew upward with increasing vigor as it penetrated better illuminated shallower waters. However, it is not clear that this biotic assemblage possessed the potential for net growth in the zone of vigorous wave action. Such potential would have to be furnished by the fenestrate bryozoans and/or the algae that are here interpreted as lime mud-forming types. The other biotic constituents either do not appear to possess frame-building qualities or are quantitatively insignificant. Elias (1948, 1949) and Elias and Condra (1957) have advocated the reef-forming importance of fenestrate bryozoans. They consider that a fenestrate meshwork could withstand wave turbulence and that it would form a wave-resistant mass when infilled with lime mud (or sparry calcite).

In the Mississippian bioherms in New Mexico evidence has not yet been noted to indicate that the core facies grew into vigorously agitated surface waters, although it may have grown upward to near the level of abrasion. Newell (1955, p. 306) notes that some lenticular reef limestones of Permian age of West Texas contain abundant fenestrate bryozoans. These are interpreted to have grown on reef knolls below wave base. The profile of the Mississippian bioherms observed in New Mexico is characteristically convex upward, and no evidence suggestive of lateral expansion of the core facies when it reached a zone of surface waters has been noted. Some surfaces of lateral planation might be expected if the core facies had grown into very shallow waters. Except for unconformities truncating the bioherms, which are clearly later than the period of biohermal growth, such erosional surfaces have not been recognized. In the grading spectrum (Lowenstam, 1950, p. 435) that exists in nature between true reef assemblages and those biotic assemblages that only form carbonate masses in quiet water environments, the writer tentatively places the fenestrate bryozoan-algal (?) assemblage of the Mississippian bioherms of the southwest somewhat below true reef assemblages. Perhaps the most critical need for detailed work on these structures is to furnish data bearing on this important genetic problem.

In conclusion, it appears that the role of fenestrate bryozoans in the formation of bioherms of Mississippian age has not been given the consideration it deserves in North

America. These organisms were long ago recognized as major faunal components of the "recifs" in the Waulsortien facies of western Europe, and their significance has been discussed at some length with regard to the Carboniferous reef knolls of England. The Waulsortien facies and the English reef knolls are of lower Carboniferous age, and essentially contemporaneous with those of southwestern United States. Elias is cited in Parkinson (1957, p. 519) as considering fenestrate bryozoans to be of major frame-building significance for the English reef knolls. Parkinson, long a student of these somewhat enigmatic masses, now favors this interpretation (1957, p. 519).

Detailed discussion of the role of fenestrate bryozoans in the formation of the bioherms in New Mexico or elsewhere is beyond the scope of this preliminary article. An excellent summary by Duncan (1957) has recently been published which critically reviews the paleoecology of bryozoa, discusses their bioherm or reef-forming potential, and summarizes the pertinent foreign and North American literature on the subject. If the present Elias-Parkinson interpretation is correct for the English reef knolls and if detailed studies confirm the writer's interpretation of the New Mexico and Texas bioherms of Mississippian age, it seems probable that fenestrate bryozoans are of greater importance in bioherm genesis than is indicated by Duncan's critical review, which was based on literature existing prior to 1956.

It is likely that the surface exposures of Mississippian bioherms of southern New Mexico will furnish needed data for many interpretive problems associated with this type of structure, not the least of which is the role of fenestrate bryozoans in carbonate build-ups. In contrast to many of the European carbonate masses of comparable age, the New Mexico bioherms are excellently exposed in three dimensions, the strata are but little deformed, and recrystallization does not appear to have obliterated much of the record of the primary faunal constituents. Unfortunately, it appears that the potentially significant part played by algae will remain speculative.

ACKNOWLEDGMENTS

The New Mexico Bureau of Mines and Mineral Resources sponsored the writer's mapping of the Sacramento Mountain escarpment from 1946 to 1950. Lowell R. Laudon and Arthur L. Bowsher were very helpful, both through their published articles and through personal discussions, as the work of these men provides the stratigraphic framework and many details regarding the bioherms of the Sacramento Mountain area. The possible significance of fenestrate bryozoans in the core facies of these bioherms was first suspected by petrographic examination of two specimens collected by Barclay Kamb and Ronald L. Shreve during the summer of 1957. Alan S. Horowitz, who assisted in subsequent field work, R. Dana Russell, and John L. Wray, all of the Denver Research Center of the Ohio Oil Company, have been helpful in discussion and manuscript review. Paul Giraudin, Jr. of the Ohio Oil Company at Wichita Falls, Texas, obtained the Texas well core and furnished geologic data regarding its stratigraphic position. John W. Harbaugh of Stanford University furnished two specimens from bioherms of northeast Oklahoma. Heinz A. Lowenstam of the California Institute of Technology has critically reviewed the manuscript.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 274–285
FIGS. 1–6, SEPTEMBER, 1958

DISTRIBUTION OF CALCIUM CARBONATE IN THE SHELF SEDIMENTS OFF EAST COAST OF INDIA[1]

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ABSTRACT

Distribution of calcium carbonate in the sediments of the continental shelf off east coast of India from Ganges to Madras has been studied with reference to the distance from the coast, configuration of the shelf, and the silt-clay component of the sediments. The shelf is covered to a very large extent by terrigenous muds. Only on the outer shelf off Madras and Kakinada-Santapilli coast, do the sediments contain more than 50 percent calcium carbonate, and these are classified as carbonate sediments. The carbonate sediments are coarse in nature and exhibit their maximum spread off Visakhapatnam where the calcium carbonate increases from 5 percent near shore to 75–80 percent at 60 fathoms depth. Beyond the continental shelf it decreases to about 10 percent. On the basis of carbonate and silt-clay contents of the sediments in relation to the shelf topography and hydrodynamical conditions, the shelf, particularly off Visakhapatnam, is divided into three zones: (1) zone of active deposition of terrigenous material extending from coast to 33½ fathoms, (2) zone of moderate deposition of terrigenous material existing between 33½ and 52½ fathoms, and (3) zone of slow or non-deposition outward of 52½ fathoms depth. The sediments in the zone of slow deposition which is also the calcareous sediments zone, are believed to have been laid down when the sea level was relatively lower than now and to have not yet been covered by the present day deposition of terrigenous material.

INTRODUCTION

With the visit to the Andhra University of Dr. E. C. La Fond in 1952–53, interest was aroused in oceanographic research, especially in the branch of marine geology. With the help of the minesweepers placed at the disposal of the University by the Govt. of India, about 400 samples of marine sediments were collected in 50 cruises organized during 1952–57. The continental shelf from "Swatch of no ground" in the north to as far as Madras in the south has been covered in a series of cruises (fig. 1). Preliminary results of studies of various aspects of the sediments, such as lithological composition, mineralogical constitution, and the bearing of the coastal rock formations on the marine sediments were reported (Mahadevan and Poornachandra Rao, 1954). The chemical composition and nature of occurrence of oolites in the sediments were also examined (Poornachandra Rao, 1955). Adequate information is, however, not available on the distribution of calcium carbonate in the sediments of the continental shelf fringing the east coast of India. A few values of calcium carbonate, as determined by Alcock in the sediments of the deeper waters of the Bay, are on record (Sewell, 1925, p. 37–38). Of a number of samples for which Alcock estimated the percentage of calcium carbonate, only one is from the shelf area, from a depth of 95 fathoms (15°56′.20 Lat. N. and 81°26′.10 Long. E.); its calcium carbonate value is 15 percent. The present studies have, therefore, been undertaken to determine the nature and distribution of calcium carbonate in the continental shelf in relation to the topography of the sea floor, influence of the river systems draining into the Bay, and other environmental factors.

[1] Manuscript received March 14, 1958.

EXPERIMENTAL PROCEDURE

Calcium carbonate content (which also includes other carbonates such as magnesium) in the sea floor sediments was determined by the author, first by a simple rapid titration method (Piper, 1947, p. 135). About one third of the same samples were later analyzed for their calcium carbonate content by the method introduced by Hutchinson and MacLennan (Piper, 1947, p. 130) which gives values of the highest order of accuracy. On the basis of the latter results, the values as obtained by the former method (which are slightly higher) were slightly modified to put the entire set of values of calcium carbonate on a uniform scale. Twenty-five samples collected off Visakhapatnam from depths greater than 40 fathoms were each separated by wet sieving through a 240 Mesh British Stand-

ard Sieve into two fractions of coarse (larger than 0.066 mm in size) and fine (less than 0.066 mm in size) silt-clay material; their percentages were computed. The coarse material was again divided into fractions of greater than 500, 500–251, 251–124, and 124–66 microns in size by use of a rotap fitted with British Standard Testing Sieves. Calcium carbonate was determined in each of these fractions as well as in the silt-clay portion in order to study the carbonate variation with grain size. The samples were also studied under the binocular as well as the petrographic microscope to determine the nature of the carbonate constituents of the sediments.

DESCRIPTION OF THE CONTINENTAL SHELF AND CARBONATE DISTRIBUTION

The continental shelf along the east coast of India is about 24 miles wide on an average except in the vicinity of the large deltas where it is either too wide or too narrow. The shelf from "Swatch of no ground" to Madras may be divided into four regional shelf units (fig. 1): (1) Swatch of no gound to Chilka Lake, (2) Chilka Lake to Kakinada, (3) Kakinada to Pennar Confluence, and (4) Pennar Confluence to Madras. Considered environmentally, the units (1) and (3) can be grouped together since they exist near river mouths. The shelf off Chilka

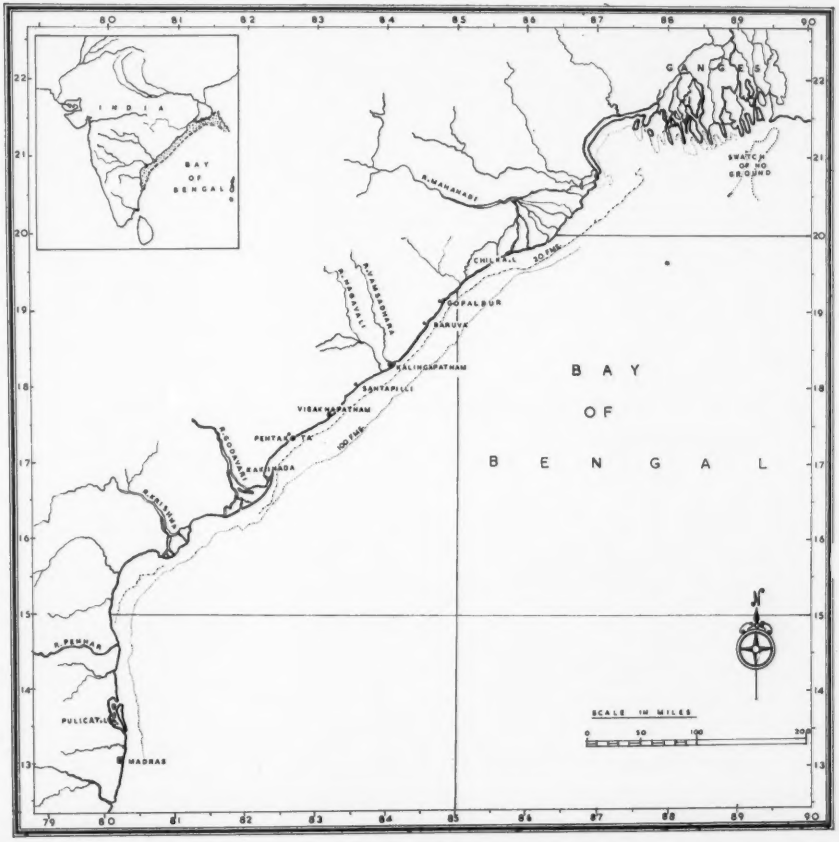

FIG. 1.—Area location map. Inset: Map of India showing the general location of the area.

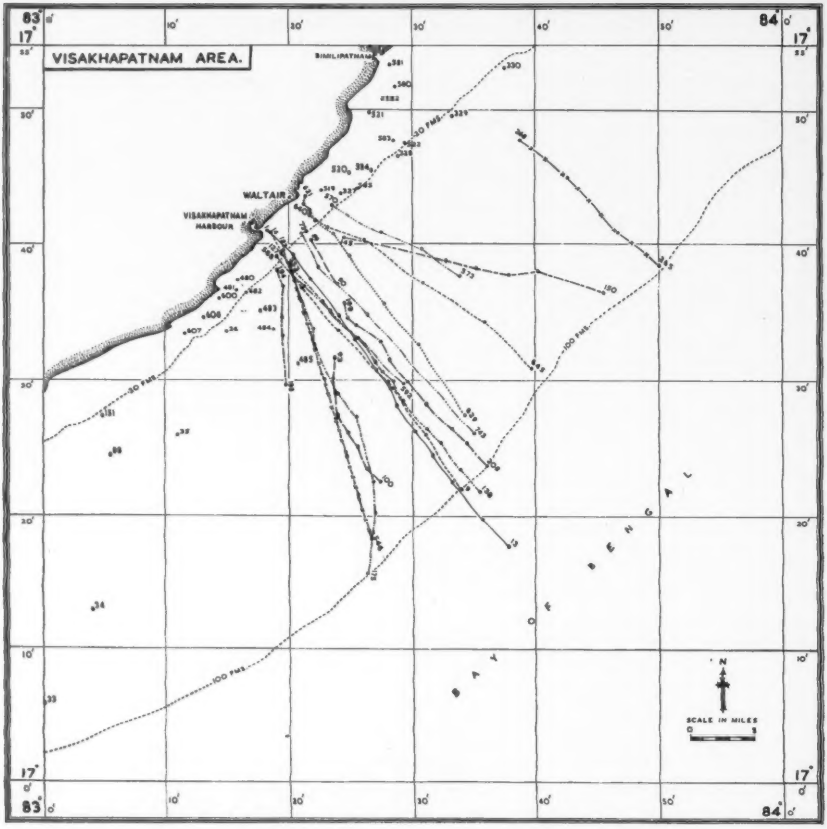

FIG. 2.—Station location map. The lines across the shelf indicate the traverses and the numbers on either end of the lines refer to Andhra University Oceanographic station numbers.

Lake-Kakinada coast is not influenced by the discharge from any major river except for a few ephemeral rivers and rivulets which are only active in the monsoon period. As might be expected, their sediment load is not considerable. This part of the shelf is subdivided into three geographical units: (a) Chilka Lake-Visakhapatnam, (b) Visakhapatnam, and (c) Visakhapatnam-Kakinada. The division has been made on the intensity of the sampling of the shelf and is purely arbitrary. The shelf off Visakhapatnam coast is chosen for examination in greater detail as a large number of samples are available from this area.

CARBONATE DISTRIBUTION OFF VISAKHAPATNAM AREA

The locations of the stations where samples of bottom materials were obtained are shown in figure 2. One hundred and twenty-five samples occupying positions along 14 traverses (which are not parallel to each other but are disposed radially with reference to Visakhapatnam) were chosen and their calcium carbonate contents were determined. From the percentage values of calcium carbonate at the corresponding station locations, contour lines were drawn at an interval of 20 percent calcium carbonate (fig. 3). The contour lines of the carbonate

roughly followed the bottom contours. The distribution against depth and distance from the coast, of calcium carbonate across the shelf off Waltair Point, Visakhapatnam and Pigeon island is presented in figure 4. Along each traverse extending out from shore to the sea, there is an increase of carbonate seaward down to near the shelf edge where it registers a sharp fall. But the rate of increase is not uniform. It rises steadily from less than 5 percent near shore to 25 percent at 40 fathoms depth and in the 40–45 fathoms area it ranges from 25 to 40 percent. In that part of the shelf of 45–50 fathoms depth it leaps abruptly to more than 55 percent. From that point onwards it maintains its upward trend until it is 75–80 percent at about 60–65 fathoms depth. On the outermost shelf in depths of 65–100 fathoms it fluctuates between 60 and 80 percent. Down the continental slope below 100 fathoms the sediments contain surprisingly low calcium carbonate content (about 10 percent), quite in agreement with Alcock's

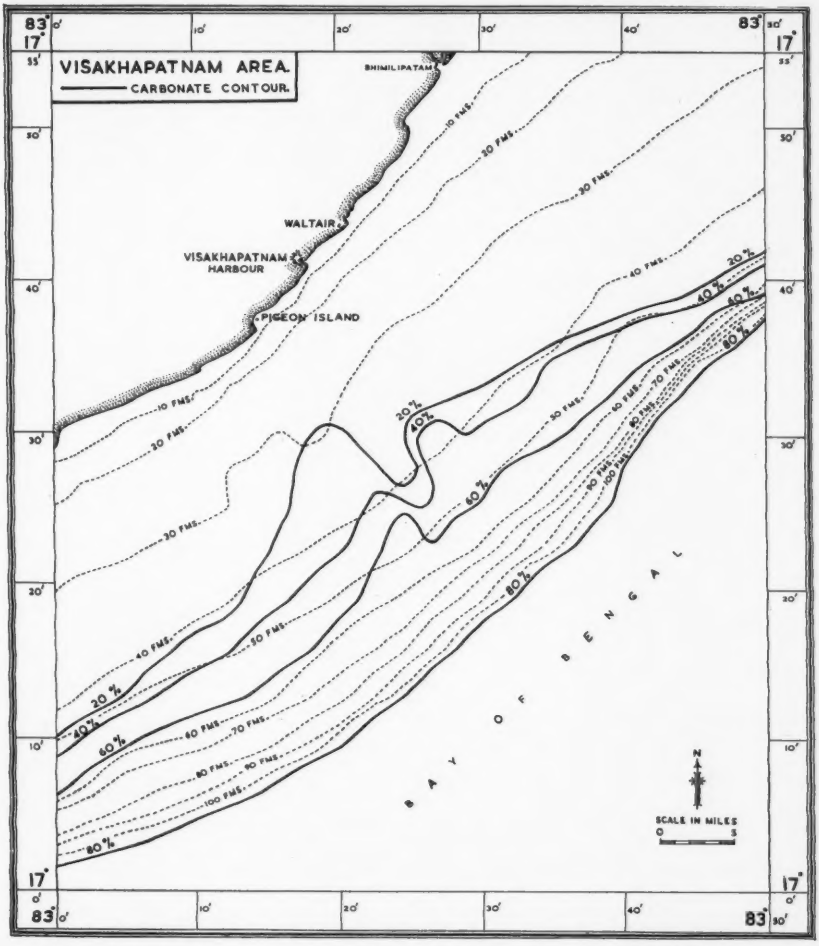

FIG. 3.—Distribution of percentage calcium carbonate in sediments off Visakhapatnam coast.

finding (Sewell, 1925, p. 38). When the silt-clay percentages and carbonate percentages are expressed as a function of depth it is found that there are two peaks of abundance of carbonate corresponding to two nadir points of silt-clay (fig. 5). The first peak of carbonate occurs in the vicinity of 60 fathoms while the second one occurs around 85 fathoms depth. In between, there is a relatively low concentration of carbonate which is also marked by an increase in the silt-clay content. When the carbonate content is analyzed in the fine fraction it is found to range from 8.5–18 percent. Even that amount of carbonate is largely due to the oolites of microscopic dimensions. Again, when the silt-clay material is removed and the coarse fraction alone is analyzed, a sharp increase in the percentage value of the carbonate is observed (table 1). The coarse fraction is subdivided into four different size groups and their carbonate contents are found to be unrelated to the size of the material except in the case of the finest fraction where a slight decrease is noticed (table 2). It appears that the carbonate percentage is dependent on the silt-clay contents of the sediments though no definite relationship can be established as yet between the two parameters. An examination of the data in Table 1 leads to the generalization that at least in the outer shelf sediments, the less the amount of fine material the higher will be the carbonate content. The few cases of deviation from this generalization may be traceable to the oolites which contain noncarbonate nuclei like quartz and felspar grains (Poornachandra Rao and Mahadevan, 1955) whose size and number alter the relationship between carbonate and silt-clay.

A visual study was made of the samples under the binocular microscope. Oolites, well rounded to ellipsoidal in shape and brown to grey in color, tests of foraminifera (pelagic as well as benthonic), shell fragments of molluscs, echinoids, bryozoans, and other organisms are the chief source of calcium carbonate. An examination under the petrological microscope had not shown particulate carbonate or carbonate minerals like calcite contributing to the concentration of calcium carbonate.

Based on the amount of calcium carbonate present, the sediments are divided into three major groups (Emery, Butcher, Gould, and Shepard, 1952):

1.—Clastic (or terrigenous) sediments with less than 50 percent calcium carbonate.

2.—Calcareous (or carbonate) sediments with more than 50 percent calcium carbonate.

3.—Intermediate mixed clastic calcareous sediments with 50 percent calcium carbonate.

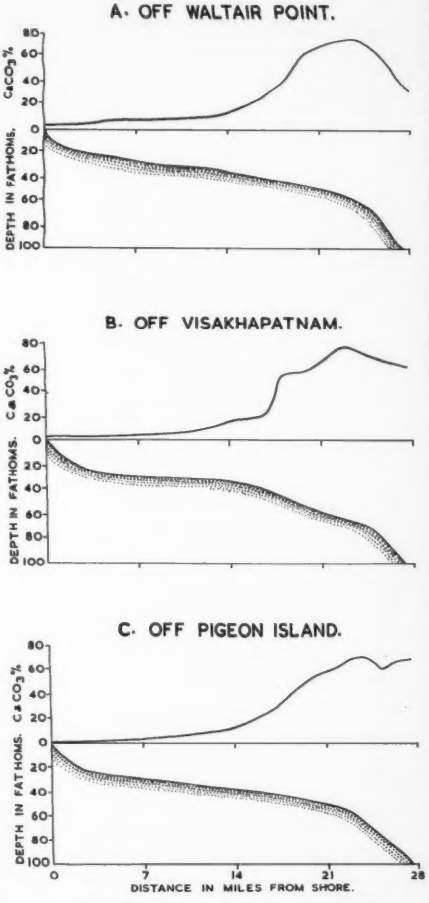

FIG. 4.—Calcium carbonate distribution in relation to depth, distance from the coast and shelf profile, off Waltair Point, Visakhapatnam and Pigeon island.

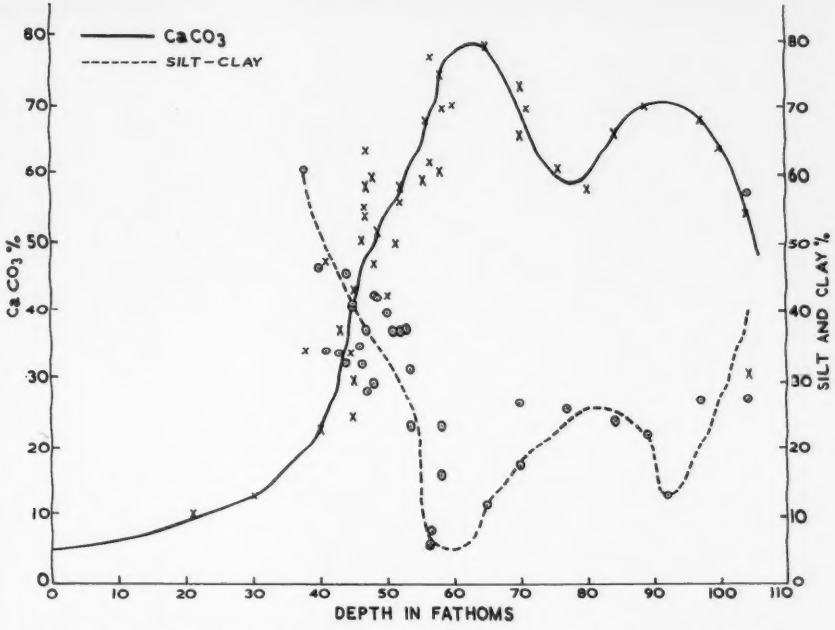

FIG. 5.—Relationship of calcium carbonate with silt-clay content and depth of the shelf.

TABLE 1.—*Distribution of calcium carbonate in the outer shelf sediments off Visakhapatnam*

S.N.	St.N.	Lat.N. in Degrees & Minutes	Long.E. in Degrees & Minutes	Depth in Fathoms	% Silt-Clay	% $CaCO_3$ in Whole	% $CaCO_3$ in Coarse Fraction	% $CaCO_3$ in Fine Fraction
1	599	17°31.2′	83°26.0′	38	61.3	34.0	76.5	—
2	131	17 30.8	83 27.7	40	46.3	22.5	62.0	8.5
3	132	17 30.1	83 28.7	43	33.9	37.0	49.5	9.0
4	95	17 28.8	83 23.7	44	32.2	33.0	43.0	—
5	592	17 29.5	83 29.8	45	40.9	24.0	33.5	8.8
6	600	17 29.5	83 27.4	45	40.7	35.0	71.0	8.8
7	548	17 22.9	83 24.9	47	38.5	63.5	83.0	15.0
8	133	17 29.3	83 28.6	47	37.4	58.0	73.5	10.3
9	314	17 31.6	83 33.9	48	29.6	59.5	82.0	—
10	96	17 27.5	83 23.8	48½	42.3	51.5	77.0	—
11	601	17 28.0	83 29.0	52	37.0	58.0	84.0	11.0
12	134	17 28.25	83 29.5	53	37.5	59.0	88.5	9.5
13	629	17 27.5	83 35.0	53½	22.8	66.0	85.0	12.5
14	97	17 26.0	83 24.8	53½	31.5	61.5	82.0	—
15	602	17 27.0	83 29.5	58	22.9	60.0	84.0	11.0
16	315	17 29.5	83 36.4	58	15.6	74.0	82.0	—
17	150	17 36.4	83 45.5	65	11.5	79.0	86.0	18.0
18	99	17 23.5	83 26.3	70	17.6	73.0	84.0	—
19	100	17 22.3	83 27.4	84	24.1	66.0	83.0	—
20	604	17 21.5	83 34.0	89	22.3	70.0	80.0	10.5
21	546	17 16.5	83 27.0	100	13.0	85.0	87.0	12.0
22	579	17 26.4	83 39.0	104	27.3	54.0	78.0	11.0
23	316	17 27.5	83 33.9	104	57.6	31.0	67.5	—
24	437	—	—	> 40	45.8	62.5	78.0	—
25	438	—	—	>100	45.3	62.5	76.5	—

TABLE 2.—*Carbonate variation with grain size in the outer shelf sediments off Visakhapatnam*

S.No.	St.No. with Depths in Fathoms in Parenthesis	% $CaCO_3$ in					
		The Whole Sample	Fraction 500 Microns in Size	Fraction 500—251 Microns in Size	Fraction 251—124 Microns in Size	Fraction 124—66 Microns in Size	Silt-Clay Fraction
1	600 (45 f)	35.0	69.0	73.0	81.5	62.0	9.8
2	548 (47 f)	63.5	87.0	85.0	83.0	73.0	15.0
3	601 (52 f)	58.0	91.0	89.0	83.0	73.0	11.0
4	134 (53 f)	59.0	92.0	83.5	85.5	86.5	9.5
5	629 (53½ f)	66.0	91.0	83.5	83.5	82.0	12.5
6	602 (58 f)	60.0	90.5	83.0	83.0	79.0	11.0
7	150 (65 f)	79.0	90.0	93.5	85.5	86.5	18.0
8	604 (89 f)	70.0	85.0	70.5	86.0	79.0	10.5
9	546 (100 f)	85.0	87.0	86.5	87.0	87.0	12.0
10	579 (104 f)	54.0	84.0	79.5	81.5	67.0	11.0

In accordance with this scheme of classification, the marine sediments of the Visakhapatnam region can be classified as clastic sediments in the zone of 0–45 fathoms and also beyond 100 fathoms and calcareous sediments in the zone of 50–100 fathoms. Taking also into account the texture of the sediments, the clastic sediments allow a further sub-division into sands, silty clays, and sand-silt-clays and similarly the calcareous sediments into calcareous organic (shells and oolites) and calcareous oolitic (oolites and shells) sediments. Shells and their fragments are mainly of molluscs and bryozoans in the former while in the latter they are mainly planktonic and benthonic foraminifera.

Table 3 brings out a comparison of the present classification of the sediments with that originally proposed by Mahadevan and Poornachandra Rao (1954, p. 27).

The two classifications are broadly identical but for the additional sand-silt-clay and the mixed clastic calcareous sediments occupying a position in between the clays and maximum shell zone of the former. Though the sand-silt-clay zone was not named as such, Mahadevan and Poornachandra Rao (1955) recognized the unusual occurrence of coarse sand which causes a break in the progressive diminution of the grain size obtaining in the sediments from shore to about 30 fathoms depth where the coarse sand begins to appear.

CARBONATE DISTRIBUTION OFF CHILKA LAKE-VISAKHAPATNAM COAST

The shelf following the Chilka Lake-Visakhapatnam Coast is uniformly wide except off Kalingapatnam and Chilka Lake mouth. The outer shelf is narrower than the inner and particularly off Santapilli, it is almost insignificant. Samples were obtained across the shelf off Chilka Lake, Gopalpur, Buruva, Kalingapatnam and Santapilli (fig. 1). In addition between Kalingapatnam and

TABLE 3.—*Classification of marine sediments off Visakhapatnam*

Zonal Classification of Mahadevan and Poornachandra Rao	Present Classification on the Basis of		% $CaCO_3$
	Composition	Texture	
Sands 0–15 fathoms	Clastic Sediments 0–45 fathoms	Sands 0–15 fathoms	0–5
Clays 15–30 fathoms		Clays 15–30 fathoms	5–15
		Sand-Silt-Clay 30–45 fathoms	15–50
Shells 30–70 fathoms	Mixed-clastic calcareous sediments (?) 45–50 fathoms.	Sand-Silt-Clay 30–45 fathoms	50
Concretions 70–100 fathoms	Calcareous Sediments 50–100 fathoms	Shelly sands	50–80
Clays beyond 100 fathoms	Clastic sediments beyond 100 fathoms	Oolitic sands	60–80
		Clays	10–15

Gopalpur samples of the bottom sediments were obtained parallel to the coast at depths of 20 and 40–50 fathoms. The carbonate distribution in this area is shown in figure 6. Off Chilka Lake, the entire shelf does not contain carbonate exceeding 5 percent. Off Gopalpur, the carbonate steadily increases from 5 percent at 10 fathoms to 27 percent at 55 fathoms and 35 percent at 85 fathoms depth. But off Baruva, it is 2 percent at 19 fathoms, 42.5 percent at 38 fathoms and only 6 percent at 74 fathoms. Beyond that it is less than 10 percent. Off Santapilli, it records a maximum of 15 percent at 45 fathoms depth.

The shelf sediments between Kalingapatnam and Gopalpur are grouped into silty sands (5–20 fathoms), medium grained sands (20–40 fathoms), shell sands (40–60 fathoms), and clays beyond 60 fathoms (Mahadevan and Subba Rao, 1955). Even the shell sands do not contain more than 50 percent carbonate. It is, therefore, obvious that the inner shelf shallower than 40 fathoms as well as the outer shelf beyond 60 fathoms depth is covered by clastic deposits. That part of the shelf in 40–60 fathoms perhaps contains mixed clastic calcareous sediments.

CARBONATE DISTRIBUTION OFF VISAKHAPATNAM-KAKINADA COAST

The shelf off Visakhapatnam-Kakinada coast is also uniformly wide, about 28 miles on average. The outer shelf deeper than 50 fathoms is only 8 miles in width and is narrower than the inner shelf just as at Visakhapatnam. Samples are available across the shelf off Pentakota and along the 17 degree parallel. A good number of representative samples are also available along 20 and 40 fathom lines. Off Pentakota the carbonate content is 19 percent at a depth of 12½ fathoms and about 10 percent beyond that station. Again at a depth of 48 fathoms it is 48 percent and at 70 fathoms, 22 percent. When these data are compared with similar information concerning the samples from 20 and 40–45 fathoms depth all along the coast, whose carbonate percentages never exceed 25 percent, it leads to the conclusion that the outer shelf in depths greater than 50 fathoms is covered towards Visakhapatnam with carbonate sediments which tend to be replaced by intermediate mixed clastic calcareous deposits and finally by a veneer of Godavari muds towards Kakinada. The inner shelf is carpeted by clastic material.

CARBONATE DISTRIBUTION OFF PENNAR-MADRAS COAST

A very narrow shelf consisting of brown and green muds derived from the rivers fringes the Kakinada-Pennar coast. The shelf begins to widen south of the Pennar Confluence and the fine muds yield place to coarse sediments. A few samples are collected off Pennar-Madras coast. Off Madras the carbonate increases from 6 percent at 9 fathoms to 69 percent at 60 fathoms and it is 25 percent at 105 fathoms depth, thus reflecting similarities with conditions prevailing off Visakhapatnam.

CARBONATE DISTRIBUTION OFF RIVER MOUTHS

The Gangetic River system enters the Bay of Bengal at its head in the northwest end while Mahanadi opens into it about 200 miles south of Ganges. These two together actively influence the shelf from "Swatch of no ground" to Chilka Lake. In the south, Godavari, Krishna and Pennar are the major rivers of importance, and their sediments settle down on the narrow shelf following the Kakinada-Pennar coast. The shelf is about 100 miles wide off Ganges and it is less than 15 miles in the case of the south Indian rivers.

Calcium carbonate in the sediments covering the shelf off these river mouths ranges from 2 to 10 percent. While the sediments off Ganges and Mahanadi contain not more than 5.5 percent of calcium carbonate, the sediments off the south Indian deltas contain 5–11 percent of it. The carbonate does not indicate any significant variation along sections across the shelf off these river mouths and along lines spreading laterally from these river confluences to considerable distances. The calcium carbonate content is thus not related either to distance from the river mouths or to the depth of the bottom. This situation compares favorably with that of Mississippi Delta area where the marginal sediments vary between 2 and 12 percent $CaCO_3$ (Shepard, '56). The relatively low occurrence of calcium carbonate in the Gangetic sediments may be due to

the extreme dilution of it by enormous quantities of detrital material poured in by Ganges and Mahanadi. The warm calcium carbonate-saturated surface waters of the Indian ocean which might contribute some carbonate, may not reach as such, as far as the head of the Bay of Bengal and therefore the possibility of carbonate precipitation is is precluded.

DISCUSSION

Murray (1910) appreciated the difference in the types of bottom deposits obtained in the northern and southern areas of the Bay of Bengal. Sewell (1925, p. 33–34) believed that on the western side of the Bay a deposit of brown mud interspersed with occasional patches of green mud can be traced all along the east coast of the great Indian Peninsula as far as the north end of Ceylon. Turning to the deeper central portion of the Bay, he draws a line from the north point of Ceylon to the northern extremity of the Andaman islands and divides the Bay into a northwestern area of terrigenous muds with low calcium carbonate content and a southeastern area of *Globigerina* ooze. He observes that the confinement of the terrigenous muds to the northwestern area is caused by the Indian and Burmese rivers emptying their silt-laden waters in this area. The latter are restricted to the coastal region by the northeast monsoon and the southerly current from October to January, while the southwest monsoon and northerly current, active from July to September, carry the waters towards the head of the Bay.

Sewell drew his conclusions mainly from observations of the deep water sediments. He did not consider in detail the shelf sediments which are not similar to those of the deeper central Bay. The few samples he obtained of the shelf material are only from near river confluences which do not contain much calcium carbonate. By collating the whole data on calcium carbonate of the shelf sediments, it is found that at Madras and between Kakinada and Chilka lake, carbonate sediments are accumulating on the outer shelf (fig. 6). They attain their greatest extent of development off Visakhapatnam. This strip of carbonate sediments disappears off Santapilli where the outer shelf in 40–100 fathoms is less than a mile wide. North of this place the outer shelf which is more than 60 fathoms deep is covered with plastic grey to green colored muds which finally grade into the river sediments of Mahanadi and Ganges. But the inner shelf between Kalingapatnam and Gopalpur is covered with sediments containing 20–30 percent carbonate unlike off Visakhapatnam. Towards south of Visakhapatnam the carbonate sediments of the outer shelf are gradually replaced by terrigenous muds which finally merge into the Godavari sediments. The exact limits of the carbonate sediment zone cannot be defined unless a very detailed and systematic sampling of the bottom material has been done.

The zone of calcareous sediments which spans the entire outer shelf off Visakhapatnam may be regarded a relict of the past or a consequence of the deposition of calcium carbonate at the present day as in the Gulf of Mexico-Caribbean Sea region, the Great Bahama Bank and many other places (Rodgers, 1957). In either case, this zone should not have been receiving terrigenous material which could dilute the concentration of calcium carbonate. The outer shelf beyond 50 fathoms contains less than 25 percent of silt-clay material. In the coarse fraction the mineral grains are only far and few in between. Poornachandra Rao (1957) made a detailed study of the oolites and the associated foraminifera occurring in this zone and came to the conclusion that the oolites were formed during a period when the sea stood at a lower level than at present. In support, he shows that the oolites are found along with shallow water forms of benthonic foraminifera like *Rotalia beccarii*, warm miliolidae, principally *Quinqueloculina, Elphidium* cf. *chrispum* which cannot thrive there under the present physico-chemical conditions. Oolites are also observed in considerable quantity in the sediments of the outer shelf off Madras even though between Kakinada and Madras the shelf is mostly covered with soft muds. North of Visakhapatnam too, as far as Gopalpur, oolites are found occurring in the coarse grained sediments wherever they exist in depths of more than 40 fathoms. From this, it is believed that the oolitic sediments covered the present outer shelf all along the coast during the period of lowered sea level and that part is now, to a large

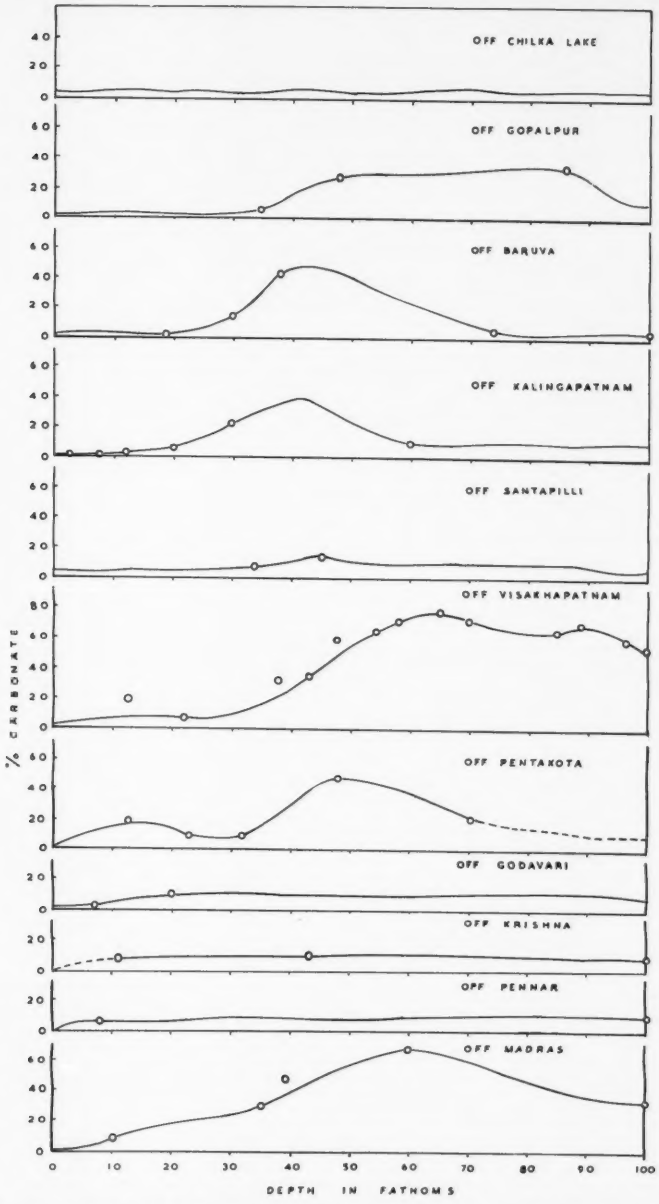

Fig. 6.—Calcium carbonate distribution across the shelf in relation to depth of bottom, off several places on the coast from Ganges to Madras.

extent, buried under the terrigenous material. Further, the coarse sand at depths of 30–40 fathoms has been suggested as representing the former beach of the lowered sea level (Poornachandra Rao, 1955). In the light of the evidence cited above, the zone of calcareous sediments may be regarded a relict of the past and its continued existence may be attributed to two factors: (1) The low carbonate river-derived sediments brought from the south or north, have not either been transported as far as this zone or may be bypassing it; it is quite possible, as Sewell suggested, that the sediments of Godavari, Krishna, and Pennar are transported away from the shelf in a northeast direction under the influence of southwest monsoon and the northerly current, whereas the northeast monsoon and the southerly current tend to confine to the shelf the sediments of Ganges and Mahanadi, thus leaving the calcareous sediment zone in between their main trends of sediment movement without receiving much terrigenous matter and (2) the Seaward transporation of detrital material from the coast due to normal processes of sedimentation has not either proceeded to this area or is bypassing it. Salinity (Ganapati and Sriramachandra Murty, 1954) and transparency (Satyanarayana Rao, 1955) observations show that less saline and high turbid waters are only restricted to within 15 miles off shore. Both the northern and southern currents attain their maximum velocities at about a distance of 15 miles from Visakhapatnam and these might carry away with them much of the sediment load, thus preventing the terrigenous matter from passing beyond. It can be corroborated by the fact that outward of this 15 miles area, that is, on the shelf outward of 30 fathoms depth, there is generally a seaward decrease of silt-clay content which constitutes 95 percent of the bottom deposits of the shelf in 20–30 fathoms.

Kukkuteswara Rao and La Fond (1954) deciphered, along the shelf profile of Visakhapatnam, conspicuous breaks in slope at 33½, 52½ and 97 fathoms. Between the coast and 33½ fathoms depth the slope is 0°5′ while it is 0°11′ between 33½ and 52½ fathoms and 0°27′ between 52½ and 97 fathoms. Mahadevan and Poornachandra Rao (1954, p. 29) showed that each break in slope of the shelf corresponds to the outward limit of a particular type of sediment, thereby linking the topography of the shelf with deposition and erosion of the sediments. Considering now the amount of calcium carbonate in relation to the silt-clay content of the sediments, the break at 33½ fathoms represents the limit of active deposition of terrigenous material, the break at 52½ fathoms the limit of moderate deposition (presuming that the coarse sand occurring in the area of 33½–52½ fathoms is not the result of present day accumulation) and finally the break at 97 fathoms represents the limit of slow or non-deposition. Thus, the shelf off Visakhapatnam can be delineated with reference to the buildup of terrigenous material into three zones: (1) a zone of active deposition from the coast to 33½ fathoms depth, consisting of sands and clays, (2) a zone of moderate deposition from 33½–52½ fathoms composed of sand-silt-clay and mixed clastic calcareous sediments, and (3) a zone of slow deposition constituted entirely by calcareous sediments.

SUMMARY AND CONCLUSIONS

The present study reveals that calcium carbonate content is influenced by silt-clay content, particularly on the outer shelf. A general relation has been found that in the marine sediments of the continental shelf the carbonate percentage varies inversely with the silt-clay content. The source of the silt-clay material is found, inter alia, in the coast and river-borne sediments. The river-derived sediments have been shown to contain very little carbonate and so too are the coastal areas. It is, therefore, logical to expect that the carbonate content increases with distance from the coast as well as from the river mouths. This expectation is justified when the outer shelf off Visakhapatnam, being situated at a distance of 20 miles from the coast and 75 and 250 miles respectively from Godavari in the south and Mahanadi confluence in the north, is found covered with carbonate sediments. Evidence has been adduced to show these carbonate sediments, oolites preponderating in them, were formed at a time of lowered sea level and they are not as yet covered by the present day deposition of terrigenous sands and muds.

ACKNOWLEDGMENTS

The author wishes to express his grateful thanks to Prof. C. Mahadevan for his keen interest in the work. The financial assistance of Council of Scientific and Industrial Research, New Delhi, is thankfully acknowledged.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 286–297
FIGS. 1–4, SEPTEMBER, 1958

BOTTOM SEDIMENTS OF SOUTHERN PART OF PYRAMID LAKE, NEVADA[1]

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ABSTRACT

The bottom sediments of the southern part of Pyramid Lake consist of fine ashy sands near the mouth of Truckee River and of diatomaceous, ostracodal, ashy, copropelic, and sapropelic silt away from the river mouth. The wet sediments are somewhat less alkaline and have stronger reducing intensity than the overlying waters. Bituminous extracts of the sediments show the presence of probable hydrocarbons, and are characterized by a fairly high content of pigmenting substances. The quantities of amino acids and humic substances in Pyramid Lake sediments are reminiscent of oligotrophic lakes in the north central part of the United States.

The term "apatotrophic" is introduced for lakes of arid regions which have alkaline waters and high total dissolved solids but relatively low C and N content, and in which the sedimentary bitumens are characterized by relatively large amounts of pigmented organic substances.

INTRODUCTION

Pyramid Lake, which lies in Washoe County, west central Nevada, about 30 miles northeast of Reno, is the largest and deepest remnant of Pleistocene Lake Lahonton (Russell, 1885, p. 56). It was selected for this brief study because it lies in an arid region, is brackish but supports life, lies in volcanic Tertiary strata and has active hot mineral springs along its eastern side, and because several geological and limnological examinations of it have already been made by other workers.

The lake bottom was sampled in August, 1955. James R. Cornell assisted in obtaining compass bearings to the sampling stations from positions on shore. Warren Toby of Nixon, Nevada, agent for the Piute Indian Reservation, and Al Jonez, Senior Technician of the Nevada Fish and Game Commission, kindly supplied boats for the work. Thomas Trelease of the Nevada Fish and Game Commission furnished limnological data. The field and laboratory work were supported by a grant-in-aid from the Graduate School, University of Minnesota. The organic substances were studied with the aid of a grant from the National Science Foundation.

[1] Manuscript received April 21, 1958.

PROCEDURES

Core samples of the lake bottom were obtained with a Phleger corer at seven stations in the southern part of the lake as shown In figure 1. Ekman dredge samples were collected northwest of Pyramid Island and between Anaho Island and the west shore. Electrical measurements of pH and Eh were made of the top and bottom of each core sample a few hours after collecting. Specimens of the wet cores were taken to the laboratory to study the moisture content, size-fractions, and organic substances. The cores were then extruded, dried, and examined with binocular microscope. Several of the wet portions were Sohxlet-extracted with a mixture of 80 percent benzene and 20 percent methanol for 8 hours; the extracts were separated by column chromatography using heptane, benzene, and pyridine + methanol as successive eluting agents. The fractions were taken up in benzene and shaken with mercury to remove sulfur. Infra-red and ultra-violet absorption spectra were run on several of the samples.

Portions of several samples were hydrolyzed with 6N hydrochloric acid for 22 hours under reflux in order to study the amino acids. After reduction and washing the hydrolyzates were taken up in iso-propyl alcohol and run on Whatman No. 1 paper in

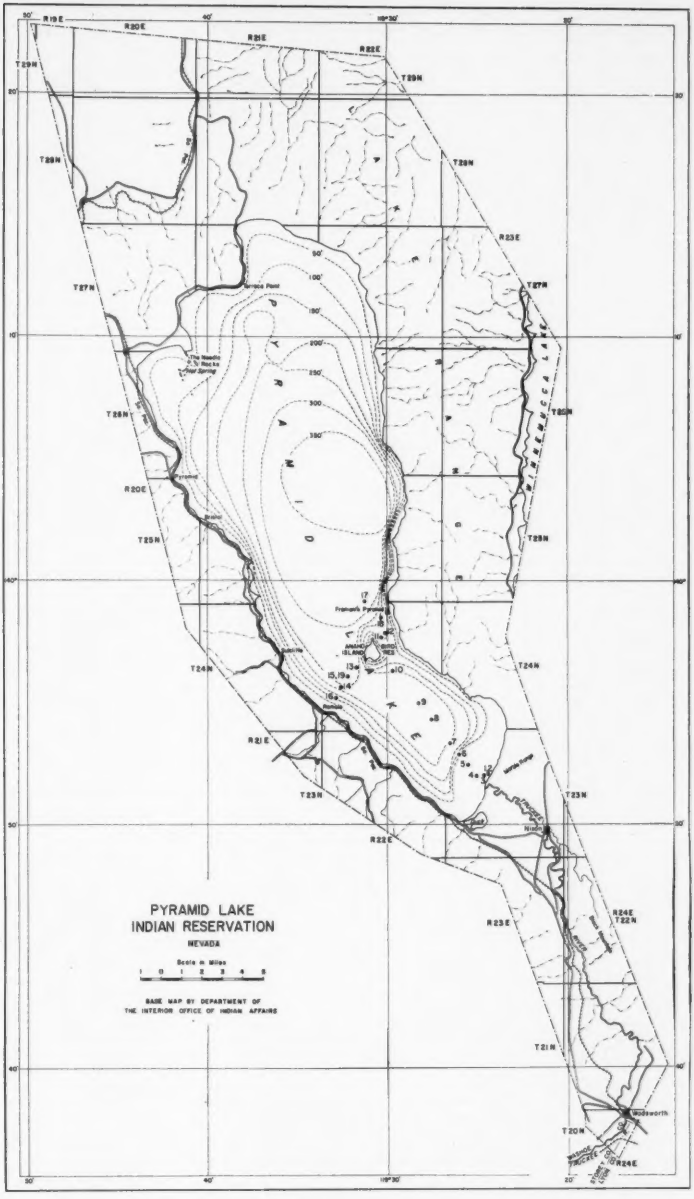

FIG. 1.—Base map of Pyramid Lake, Nevada, showing sampling stations. Bathymetric contours after Russell (1885).

butanol: acetic acid: water, phenol: water, and in collidine. The sugars were separated by hydrolyzing in the cold with 70 percent H_2SO_4 neutralizing, refluxing, reducing, and running the concentrate on paper in butanol: acetic acid: water (40:10:50).

RESULTS

Geologic Summary.—Pyramid Lake is approximately 30 miles long, up to 12 miles wide, and occupies about 800 square miles (Fairbanks, 1901); the area of greater depths is a few miles north of Anaho Island where it is more than 300 feet deep. The lake lies in a complex of later Tertiary volcanic rocks representing the Esmeralda Formation (Turner, 1900). Faulted and metamorphosed Mesozoic(?) carbonate rocks are exposed in "Marble Buttes" near the southern end of the lake. The basin is bounded by one or more faults on the eastern side, (Hutchinson, 1957, p.18) along which hot springs or gas vents issue forth. Good examples of such a hot spring occur on the western side of Pyramid Island.

Calcareous tufa has accumulated in large quantities along the beaches and cliffs of the lake. This material apparently has been deposited directly from the lake water as a result of the drainage into the lake of carbonate-charged surface water or from sublacustral hot spring sources.

Several pronounced changes of lake level have been dated by C^{14} methods on the basis of tufa samples collected from the various terrace levels of Pyramid Lake (Broeker and Orr, 1956). The C^{14} data suggest that tufa formation in Pyramid Lake first began within the last 30,000 years and that the lake has been mainly shrinking during the last 11,700 years, following a minimum about 12,500 years ago. There is, however, a level approximately 100 feet above the present lake surface below which the tufa is fresh-looking. The 100 foot level suggests a period of stillstand of considerable duration.

Morphometry of lake basin.—The Pyramid Lake basin is roughly wedge-shaped in east-west cross section with a steep eastern slope caused by marginal faulting on the flanks of the Lake Range. The western slope is relatively more gentle with slopes of from 20 to 150 feet per mile. The southern end of the lake has a steep foreset delta slope off the mouth of Truckee River; the northern part of the lake basin slopes very gently 20–25 feet per mile, as is characteristic of several of the Lahonton basins. The middle part of Pyramid basin occupying 30–50 square miles is fairly flat, a depositional feature.

Terraces mark the former level of Lake Lahonton around Pyramid Lake (Russell, 1885, p. 104–105). The highest of these is "Lohonton Beach" which is approximately 570 feet above Pyramid Lake level. Other terraces are the "Lithoid Terrace" ± 540 feet, "Dendritic Terrace", 320 feet; and "Thinolite Terrace" 150 feet. The Thinolite Terrace occurs at the top of the "unweathered" tufa. The lower three terraces are named from the variety of tufa prominent in each.

Limnologic studies emphasizing trophic characteristics and annual heat budget were made by Hutchinson (1937).

According to Jonez (1955) Pyramid Lake has recently been declining about 3 feet per year.

Composition of water.—The lake water is brackish as indicated by the analyses given in table 1, but the concentration of salts is low enough to permit several species of freshwater fish to occur and to allow consumption by cattle. Precipitation of native sulfur from the lake water is occurring, probably because of the activity of sulfate reducing bacteria.

Water movements and thermal stratification.—The limnologic work of Jonez (1955) has lead him to the following conclusions concerning major circulation of water in Pyramid Lake: warmer surface water occurs to relatively greater depths at the south end than at the north end of the lake owing to mixing action of prevailing northerly winds; the influence of silt-laden Truckee River entering at the south results in a circulation of the deep water from south to north displacing warmer water at the north end of the lake which moves southward at the surface.

The lake is thermally stratified, the thermocline in the Fall occurring from about 60 to 90 feet. Bathythermograph records are shown in figure 2. Despite the brackishness of the water and high evaporation rate, no chemical stratification has been found in the lake (table 1).

TABLE 1.—*Composition of water of Pyramid Lake, ppm*

	A	B	C	D	E
Silica	2.6	2.1	2.8	41.2	20.0
Iron	0.03	0.04	0.03		
Calcium	7.5	8.0	8.5	17.9	17.9
Magnesium	111.0	111.0	111.0	80.0	80.0
Sodium	1570.0	1540.0	1540.0	1173.1	1181.7
Potassium	128.0	160.0	160.0	76.6	72.3
Carbonate	285.0	280.0	325.0	499.8	492.1
Bicarbonate	830.0	830.0	745.0		
Sulfate	263.0	265.0	265.0	180.3	185.0
Chloride	1920.0	1920.0	1920.0	1429.8	1434.2
Fluoride	2.9	2.9	2.9		
Nitrate	1.8	1.8	2.2		
Boron	12.0	12.0	12.0		
Σ dissolved solids ppm	4700.0	4700.0	4700.0	3498.7	3483.7
Σ dissolved solids tons/acre foot	6.39	6.39	6.39		
$CaCO_3$ hardness	478.0	476.0	478.0		
Specific conductance microohms @25° C	7610.0	7660.0	7660.0		
pH	9.4	9.2	9.0		

A 10 feet from surface, February 13, 1951, Pyramid Lake, U.S.G.S., Salt Lake City Lab. #6139.
B 100 feet.
C 195 feet.
D 1 foot below surface, north of Anaho Island, August, 1882 (Russell, 1885).
E 350 feet.

Biology.—No formal studies of the biota have been made, but included are melosiroid, campylodiscoid and naviculoid Diatomacea in large numbers, *Chara* sp., *Cladophora* sp., *Potamogeton pectinatus*, Myxophytic algae, *Grantia* sp., *Osphranticum* sp., *Daphnia pulex parapulex*, *Ceriodaphnia laticaudata*, *Limnocythere reticulata*, diaptomid larva, midge larvae, and 8 species of fish, including *Chasmistes cujus*, an endemic Lahonton species found living nowhere else in the world. Anaho Island is a famous pelican rookery. Hutchinson (1937, p. 67) identified *Pyrgulopsis nevadensis* (Stearns) and *Paraphalyx effusa* (Lea) in the lake sediments. *Limnocythere* spp. and *Candona* spp. are very abundant in the core samples. The raised Lahonton lake sediments on the western side of the lake contain *Candona* spp. *Limnocythere* sp., and *Cypricercus?* sp.

Description of bottom deposits (Table 2).—No cores were recovered in the sandy foreset delta beds near the mouth of Truckee River, but light gray silty sand mixed with organic matter stuck to the core barrel. At Station 4 the bottomset delta beds in 160 feet of water consist of dark gray diatomaceous silt or silty diatomite, drying in air to pale yellowish and brownish gray, grading downward into fine-grained angular micaceous feldspathic sand. Diatoms, cladocerans, and ostracodes are abundant in the upper 12 inches but almost absent in the lower 8 inches of the core. Station 5 also encountered the bottomset silts, but they are calcareous and volcanic glass shards occur abundantly. The bottom sediments at stations 6–9, despite the variation in depth of water, are rather uniformly dark gray, diatomaceous, ashy, ostracodal silt. Core No. 9 is less calcareous and contains littoral gastropods not found in the deeper waters. Off the southern end of Anaho Island (Station 10) there is dark gray, silty and sandy, diatomaceous, ostracodal clay with naviculoid diatoms, which dries to a pale brownish gray. The sand is derived from the rhyolitic tuffaceous agglomerates that make up Anaho Island.

Core recovery was poor between Anaho Island and the mainland. The bottom is sandy in some places and covered with boulders and sticky green clay in others. A core in 235 feet of water west of Anaho Island consisted of dark gray to black very diatomaceous slity clay drying pale brownish gray and emitting a slight H_2S odor in the

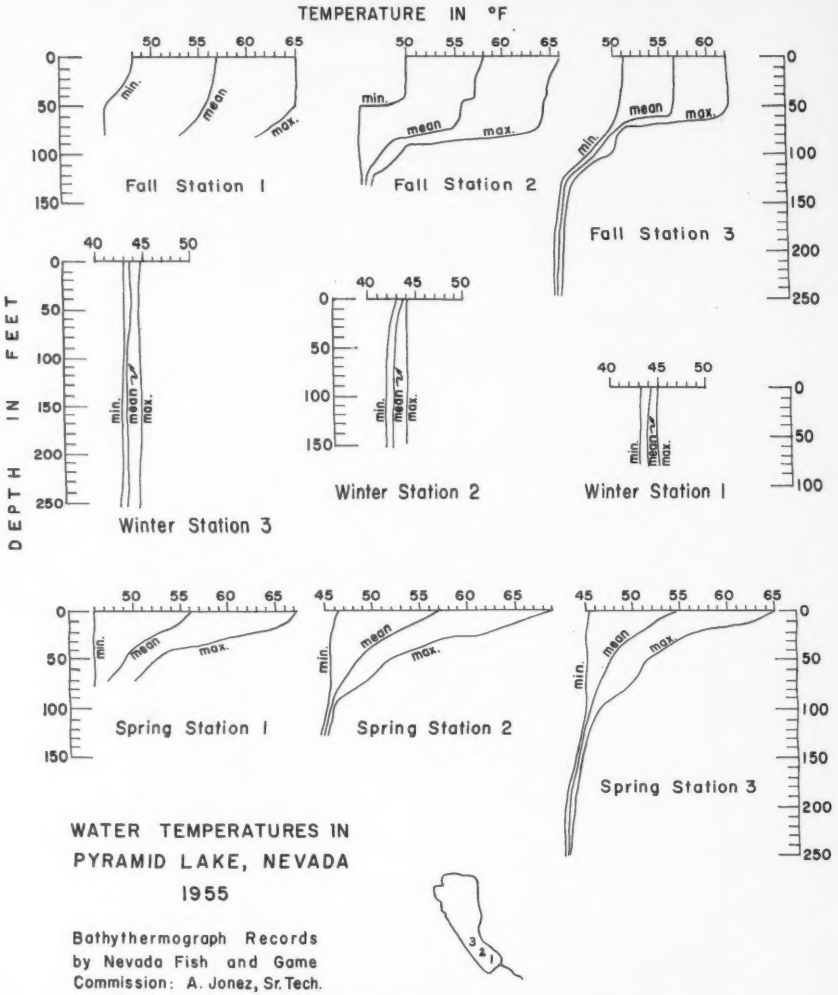

FIG. 2.—Water temperatures in Pyramid Lake for 1955. Courtesy of Nevada Fish and Game Commission, A. Jonez, Senior Technician.

fresh state.

All of the bottom samples gave strong H_2S odors when brought to room temperature although little or none was detected immediately after sampling.

The samples north and west of Anaho Island in 300 feet or more of water were obtained with an Ekman dredge. They consist of black diatomaceous clay and silt, drying pale brownish gray.

Along the southwestern shelf of the lake the bottom sediments are sublittoral gray sand that could not be cored but which was collected in several Ekman dredge sound-

TABLE 2.—*Core data, Pyramid Lake*

Station	Depth (feet)	Length of core (inches)	Description of Core (colors are those of dried cores)	Remarks
1	109		No recovery, sand and silt	
2	110		No recovery, sand and silt	
3	80		No recovery, dark colored silt	
4	159	25	Upper 12 inches, pale yellowish, brownish gray, diatomaceous silt or silty diatomite; cocconeoids, melosiroids, coscinodiscoids, *Limnocythere*, *Candona*, cladocerans Lower 8 inches, pale reddish, grayish brown, feldspathic silt; and fine grained angular micaceous sand; few ostracodes and carbonized plant fragments	
5	179	40	Pale brownish gray, calcareous, tuffaceous, diatomaceous silt; tiny volcanic glass shards very abundant; common ostracodes (mainly *Candona* spp.,) and cladocerans, few fish bone fragments	pH of water above core 8.5 (field). Moisture, top of core, 70%; bottom, 64%
6	199	27	Pale brownish gray, calcareous, very diatomaceous (melosiroid[s]) silty clay and clayey silt; volcanic glass shards; scattered ostracodes, cladocerans and gastropods in upper part, becoming more abundant in lower part; *Candona* sp., *Limnocythere* sp.	pH of water above core 8.5–9 (field)
7	195	29.5	Pale brownish gray, slightly calcareous, very diatomaceous (melosiroids) silty clay and clayey silt; *Candona* cf. *candida*, *Limnocythere* sp., cladocerans	Surface temperature 22° C. Bottom water temperature 10.2° C. Bottom core temperature 13° C. Moisture, top of core 68%. Moisture, bottom of core 61%
8	174	32	Pale brownish gray, very diatomaceous, finely sandy, micaceous, ostracodal silt or diatomite; *Limnocythere* sp. very abundant at some levels; melosiroid and campylodiscoid diatoms	Chemical analysis of bottom of core by University of Minnesota Mines Experiment Station, V. C. Bye, analyst, as follows: Fe_2O_3 10.45%, FeO 0.69%, P_2O_5 0.46%, SiO_2 49.08%, MnO_2 0.48%, Al_2O_3 14.28%, TiO_2 0.33%, CaO 7.45%, MgO 2.45%, Na_2O 2.67%, K_2O 2.34%, C 0.92%, S 0.95%, H_2O 4.12%, CO_2 3.32%
9	98	38	Pale yellowish gray, ashy, slightly calcareous diatomaceous, ostracodal clay; lower 2 inches fine grained gray sand; *Candona* sp., *Limnocythere* sp., *Cypricercus* sp. with weakly reticulate surface, cladocerans, small gastropods in lower part	
10	64	22	Pale brownish gray, silty, and finely sandy, calcareous micaceous, diatomaceous, ostracodal clay; *Candona* cf. *caudata*, *Limnocythere* sp., cladocerans, naviculoid diatoms	Moisture, top of core 51%. Moisture bottom of core 77%
11	180		No recovery, sticky green and black clay with fragments of calcareous tufa, bottom solid	
12	150		No recovery, sticky green clay, bottom solid	
13	235	30	Pale brownish gray, micaceous, calcareous, very diatomaceous, silty, ostracodal clay or diatomite; melosiroids, *Candona* sp., *Limnocythere* sp.	Slight odor H_2S in fresh core; pH bottom water 8.5
14	100		No recovery, black sapropelic ooze and sand on core barrel	
15	170		No recovery, sand	Considerable blue green algae and fragments of pondweeds floating near Romolo
16	138		No recovery, sand	
17	315		Edman dredge sample, black sapropelic ooze	pH 8.5 at bottom; noticeable H_2S odor
18	305		Edman dredge sample, black sapropelic ooze	
19	195		Edman dredge sample, sand and black sapropelic ooze	H_2S odor

ings. That shore of the lake has a beach composed of quartz sand and many empty gastropod shells which form coquinites in some places.

Moisture content.—The relatively low content of moisture for fresh sediments that characterizes these bottom deposits results from their siltiness. In Minnesota lake sediments (Swain, 1956, p. 630) it was found that highly organic sediments (copropel and sapropel) contain the most moisture, 85 percent or more, and that additions of sand and silt to form as much as half the sediment caused reduction in moisture content to below 80 percent. Shrinkage, however, is much greater in the organic than in the silty sediments, and the resulting porosity has an inverse relationship to shrinkage. The Pyramid Lake silty sediments show relatively little shrinkage.

An interesting example of the relationship between particle size and moisture content is found in core No. 10 where the low moisture in the top of the core reflects the coarser texture of that part of the core (table 3).

Mechanical Analyses.—Size fractions of the top and bottom of each core were obtained by wet sieving to obtain the sizes > 2 mm (granules and pebbles) and 2–.074 mm (sand). The sizes > .074 mm (silt and clay) were obtained by collecting the fraction on a bacteriological filter candle and

TABLE 3.—*Some properties of Pyramid Lake sediments*

Station	Position in core	pH	Eh	%>.074mm, <2 mm	% <.074 mm.	Description of fraction	Remarks
4	Bottom water	9.1	275				
4	Top of core	8.9	275	3.04	96.96	Clayey peaty silt, very micaceous	
4	Bottom of core	8.85	+173	7.93	92.07	Peaty, micaceous silty clay and silt, many plant fragments, ostracodes, other arthropod parts	
5	Bottom water	9.0	+167				
5	Top of core	8.8	+ 77	3.10	96.90	Fibrous clayey silt; ostracodes, etc.	H_2S odor
5	Bottom of core	8.9	+197	4.22	95.78	Micaceous ostracodal clay, containing brown spherical arthropod? egg cases, carbonized plant fragments	No H_2S
6	Bottom water	9.1	+309				
6	Top of core	8.6	+149 U[1]	4.27	95.73	Clay with brown arthropod? egg cases, ostracodes, diatoms	H_2S odor
6	Bottom of core	8.7	+179	5.25	94.95	Diatomaceous, ostracodal silty clay	No H_2S
7	Bottom water	9.1	+287 U	4.78	95.22	Gray ostracodal, cladoceran silty clay and silt	Slight H_2S odor
7	Top of core	8.75	+129 U	30.28	69.72	Ostracodal gray clayey silt	
7	Bottom of core	9.05	+173	n.d.	n.d.	Lost	
8	Bottom water	9.0	+287				
8	Top of core	9.1	+257	12.89	87.11	Ostracodal clay, egg cases	
8	Bottom of core	8.95	+239	11.19	88.81	Ostracodal, silty micaceous clay; plant fragments	Slight H_2S odor
9	No water						
9	Top of core	8.9	+245	20.55	79.45	Coarse ostracodal, micaceous ashy silt	Slight H_2S odor
9	Bottom of core	8.85	+239	34.69	65.19	Coarse ostracodal, ashy coquina	Slight H_2S odor 0.11% >2 mm
10	Bottom water	9.2	+309				
10	Top of core	9.05	+275	38.13	61.27	Lost	0.6% >2 mm
10	Bottom of core	9.0	+281	7.87	92.13	Coquina of *Candona* and *Limnocythere*, containing clayey aggregates and egg cases; tufa balls; shells are large and robust	
13	Bottom	9.1	+323				
13	Top of core	8.65	+181	5.89	94.11	Ostracodal clay	
13	Bottom of core	8.85	+227	4.35	95.65	Ostracodal clay	

[1] U, unsteady Eh readings.

weighing. Sizes coarser than granules are rare in these samples and mainly comprise gastropod shells and fragments of tufa. The increase in the sand fraction in the bottom of core No. 7 and the decrease in the bottom of core No. 10 probably can be correlated with a past decrease and increase, respectively, of lake level and suggest that sedimentation in Pyramid Lake has been relatively slow.

Hydrogen ion concentration (*table 3*).—The pH values of Pyramid Lake water average 9.2 at the surface and 9.1 at the bottom. The top and bottom of each core have pH values averaging 8.9. The alkaline values result from the quantities of hydroxyl ions from the sodium and potassium salts. The drop in pH from the water into the lake sediment may result in part from: (1) the production of H_2S by decaying organic matter; (2) the acidity of the tuffaceous sedimentary material; or (3) the filtration effect across the mud-water interface, such as occurs in the mud cake on the walls of rotary drill holes (Doll, 1950, p. 161).

Oxidation-reduction potential (*table 3*).—The Eh values of the bottom waters of Pyramid Lake were weakly oxidizing whereas the sediments in large part were weakly to moderately reducing. A few of the Eh readings in the sediment were weakly oxidizing. The presence of hydrogen sulfide accounts for the reduced state of the sediments. Its rapid evolution from the sediments may account for the oxidizing values of some of the samples and their poorly poised condition.

Radioactivity.—Total radioactivity of two samples of Pyramid Lake sediment was obtained using a scalar Geiger counter. Radioactivity of the top of core No. 11 was 3.60 cpm while that of the top of core No. 11 was 4.67 cpm above background, representing a U_3O_8 equivalent of $< .001$ percent.

Bitumens.—The bitumen or lipoid content of four of the core samples is given in table 4. The values are intermediate between the high ones of peat (65,000 ppm) and the low ones of oligotrophic lakes (300 ppm) and are of the same order of magnitude as of a typical eutrophic Minnesota lake (8350 ppm). Sulfur was present in all the bituminous extracts.

The chromatographic fractions eluted from the columns of activated alumina with *n-heptane* (table 5a) are colorless oil, mainly saturated hydrocarbons, and form an aver-

TABLE 4.—*Bituminous extracts of Pyramid Lake sediments; values before correction for S are in parentheses*

Station	Position in core	Wt. of extract (g)	Ppm of sample	Description of extract
5	Top 12 inches	.2167	(8200) 8188	Dark brown to black tar; tarry odor; bladed and fibrous pale yellow sulfur crystals 0.14% S
7	Top 4 inches	.1231	(6800) 6120	Brown earthy wax; long needlelike yellow crystals of sulfur; slight reddish fluorescence 9.6% S
10	General	.1523	(5400) 4990	Light brown porous earthy wax; few tiny bladed iridescent crystals 7.6% S
13	Top 6 inches	.1257	(5900) 5204	Brownish yellow tar with yellow orthorhombic crystals of sulfur 11.8% S
Average bitumens ppm of sample			6575	

age of 72 percent of the total hydrocarbons.

The fractions eluted with *benzene* (table 5b), taken to represent the aromatic hydrocarbons are also colorless oil, averaging 28 percent of the hydrocarbons. An ultra-violet absorption spectrum of one of the benzene fractions (fig. 3) shows peaks at 264, 270, and 280 mμ.

The fractions eluted with *pyridine* and *methanol* are yellowish brown wax, containing red, orange, and purple pigmenting substances, probably of tetraterpene or carotenoid nature. The pink and red fluorescence of the solutions during chromatography is suggestive of chlorophyll degradation products. The bulk of pyridine + methanol

TABLE 5.—*Chromatographic analyses of bituminous extracts from Pyramid Lake*

No.	Position in section	Wt. of fraction	Description of fraction	Ppm of sample	% of bitumens	% of non-polar bit.	% of hydrocarb.	Sulfur	Remarks
			a. Heptane chromatographic fractions (corrected for sulfur content)						
5	top 12 inches	.0224	Colorless oil, frosty white substance, trace greenish white fluorescence; lathlike crystals	852	12	45.8	82.3	Yes	Heptane cut dark, elution clear
7	top 4 inches	.0038	Mainly bladed and globular sulfur crystals; colorless oil, frosty white substance	210	3.4	26.3	65.5	Yes	Pink fluorescence in filter aid; none in alumina; a brown front developed in upper half of column
10	general	.0015	Mainly lathlike sulfur crystals, colorless bladed crystals, frosty white substance	53	1.0	16.3	45.4	Yes	Heptane cut colored, eluant clear
13	top 6 inches	.0024	Colorless oil, frosty white substance	114	2.2	16	96	Yes	Pink fluorescence in filter aid; developed brown zone in alumina
			b. Benzene chromatographic fractions (corrected for sulfur content)						
5	top 12 inches	.0048	Crystal laths of sulfur colorless oil, frosty white substance	182	2.6	9.8	17.6	Yes	Benzene cut dark, eluant yellowish fluorescent
7	top 4 inches	.0020	Crystal laths of sulfur; colorless oil with weak greenish yellow fluorescence, colorless crystals, frosty white substance	110	1.8	13.9	34.5	Yes	No fluorescence in benzene in column
10	general	.0018	Mainly lathlike crystals and globules of sulfur	64	1.3	19.5	54.5	Yes	Cut colored, eluant clear, no fluorescence
13	top 6 inches	.0001	Crystals of sulfur, trace frosty white substance	4.7	.09	19.5	4.0	Yes	
			c. Pydieine+methanol chromatographic fractions (corrected for sulfur content)						
5	top 12 inches	.0217	Yellowish brown wax	825	11.7	44.4	77	Yes	Pyridine cut very dark, elution also dark, purplish red color disseminated in alumina slightly fluorescent
7	top 4 inches	.0086	Yellowish brown wax, frosty white substance	475	7.7	59.7	87	Yes	Pyridine front had red-orange fluorescence
10	general	.0059	Yellow-tan to brown wax, frosty white substance	209	4.2	64	93	Yes	Pyridine cut light colored, elution dark, good purplish red band in alumina, slightly fluorescent
13			Yellowish-green wax, frosty white substance	594	11.2	83.3	86		

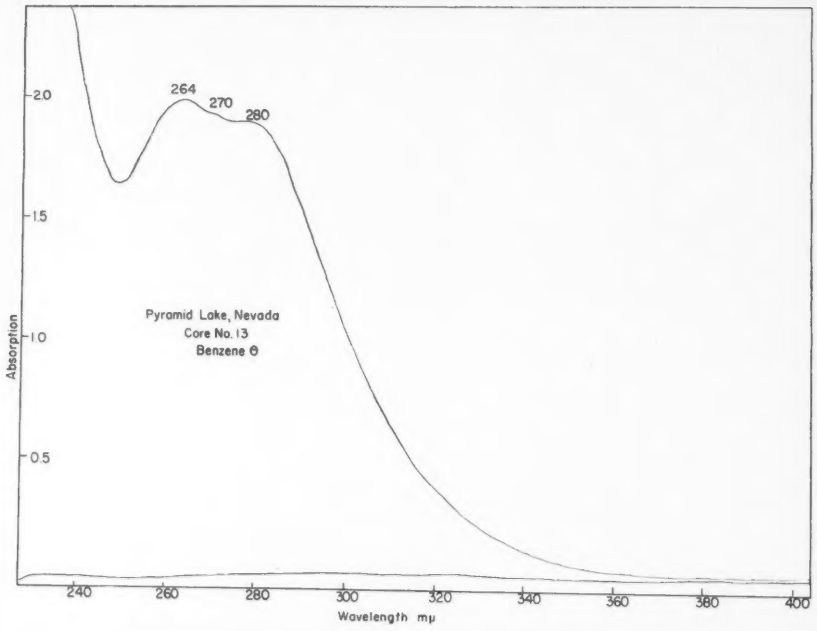

FIG. 3.—Ultraviolet absorption spectrum of fraction of bitumens from core No. 13 eluted from column of activated alumina with benzene, dried, and taken up in n-heptane.

fraction is taken to represent the asphaltenes.

The polar compounds remaining adsorbed on the alumina (table 5c) comprise pigments carbohydrates, mineral substances, and others. They form a generally higher percentage of the Pyramid lipoids than in a variety of other lakes (Swain, 1956, p. 643).

Amino acids.—The amino acids separated from two of the Pyramid Lake core samples by HCl hydrolysis are listed in table 6. The quantities of amino acids are indicative of a lake of late oligotrophic or early eutrophic nature by comparison with lake deposits elsewhere. In Minnesota lakes the quantities range from 0.1×10^{-4} gm/gm in early eutrophic marl to 50×10^{-4} gm/gm in late eutrophic and dystrophic peat (Swain, Blumentals and Millers, 1959).

Compared to Minnesota lakes, similar or somewhat larger amounts of amino acids were found by Kleerekoper (1957, p. 123) in lake sediments of southern Ontario.

Carbohydrate degradation products.—Chromatograms of sulfuric acid hydrolyzates of Pyramid Lake samples show the presence of substances of high Rf value, probably derived from carbohydrates. These are referred to as humic acid constituents (Swain, Blumentals and Prokopovich, 1958; Swain, 1959)

TABLE 6.—*Amino acid content of Pyramid Lake core samples (parts per 10,000 of wet sample)*

Amino acid	Core No. 17	Core No. 18
Aspartic acid+glycine	1.62	2.13
Glutamic acid+threonine	0.05	0.46
Alanine	0.26	0.12
Valine	0.36	0.43
Leucine	0.01	0.06
Histidine	0.12	0.01
Total	2.42	3.19

Chromatographic analyses by A. Blumentals and R. Millers.

FIG. 4.—Paper chromatograms of sulfuric acid hydrolyzates of bottom samples from Pyramid and other lakes and Devonian black shale. Solvent, butanol:acetic acid:water (4:1:5), stain, aniline and phthalic acid. 1, Glucose and arabinose. 2, Pyramid Lake, Station 18, *Rf* 0.345, plus additional spot. 3, Cedar Creek Bog, Minnesota, 41–42 feet, *Rf* 0.40. 4, Dismal Swamp, Virginia, 4–5 feet, *Rf* 0.37. Prior Lake, Minnesota, *Rf* 0.40. 6, Glucose and arabinose. 7, Big Island peat, Lake Minnetonka, Minnesota, *Rf* 0.41. 8, Sapropel, Prior Lake, Minnesota, *Rf* 0.41. 9, Copropel, Prior Lake, Minnesota, *Rf* 0.38, plus additional spot. 10, Pyramid Lake, Station 18, *Rf* 0.39, plus additional spot. 11, Glucose and arabinose. 12, Copropel, Prior Lake, Minnesota, *Rf* 0.45. 13, Devonian Marcellus black shale, Huntingdon County, Pennsylvania, *Rf* 0.43. 14, Pyramid Lake, Station 18, *Rf* 0.355 plus additional spot. 15, Catahoula Lake, Louisiana, *Rf* 0.398. Abbreviations: brn, brown fluorescence after staining with aniline, rb, reddish brown fluorescence; g, golden fluorescence; pg, pale golden fluorescence. Concentration of acid used for hydrolysis was 1N except in No. 5 in which it was 72 percent. Ultraviolet absorption spectra of the substances from the lake and rock samples having *Rf* values of 0.35 to 0.40 suggest pyridine compounds.

because of their presence in hydrolyzates of peat humic acid as well as many kinds of lake samples, bituminous rocks, and cooking starch. They are characterized by Rf values of about 0.35–0.4, golden fluorescence after staining with aniline and phthalic acid, and ultra-violet absorption peaks at 264, 268 and 272 mμ. These peaks and the general shape of the absorption curve strongly suggest one or more of the pyridines, as 3-chloropyridine (API ultraviolet absorption spectra). The principal source of the pyridines is coal tar. The humic acid constituents of the Pyramid Lake sediments are considerably less than in equal quantities of Minnesota lake sediments and Devonian black shale used for comparison but are approximately equal in amount to Dismal Swamp (Virginia) peat and Catahoula Lake (Louisiana) organic silt. Comparative chromatograms are shown in figure 4.

CONCLUSIONS

The typical bottom sediments away from Truckee River delta in southern Pyramid Lake are dark gray, ashy, diatomaceous,

TABLE 7.—*Comparison of carbon and nitrogen content of lake sediments*

Lake	Sediment	Total diss. solids in lake water	% C	% N
Oligotrophic				
Superior, Minn.	Silty clay	73	2.92	<0.5
Rainy, Minn.	Silty clay	—	3.09	<0.5
Lake of the Woods, Minn.	Copropelic silt	—	5.73	<0.5
Burntside, Minn.	Silty copropelic clay	25	10.56	0.92
Kabekona, Minn.	Marl	—	5.79	<0.5
Eutrophic				
Minnetonka, Minn.	Copropel	155	11.25	—
Kabetogama, Minn.	Copropelic silt	92	11.35	1.39
Prior, Minn.	Marly copropel	135	12.54	1.16
Cedar, Wright Co., Minn.	Sapropelic copropel	160	14.03	1.42
Johanna, Minn.	Copropel	145	16.08	1.90
West Rush, Minn.	Copropel	225	16.24	1.81
Pelican, Minn.	Silty copropel	50	24.12	2.15
Cedar Creek Marl, Minn.	Copropelic marl	—	15.13	0.75
Cedar Creek peat, Minn.	Copropelic peat	235	48.01	3.36
Apatotrophic				
Pyramid, Nev.	Copropelic silt	4700	2.67	<0.5

ostracodal silts. On drying the sediments undergo oxidation of the iron oxides and sulfides to form buff coloration. The ostracodes perhaps are no longer living or are quite rare in the lake; in this respect they are similar to the gastropods whose shells make coquinas along the shore and are evidently totally extinct. The gastropod shells generally are rare in the basinal sediments. The silty and finely sandy nature of the cores results in moisture content ranging between 51 and 77 percent, rather low figures for lake sediments, pointing to the relatively low organic productivity of the lake. Shrinkage is low in the Pyramid Lake sediments which reflects their porous character. The high alkalinity of the lake waters is the major factor in the reduced state of the sediments and the negative Eh values. In this respect the lake is very different from the typical one of low productivity which is low in dissolved solids.

The total lipoids, hydrocarbon, and asphalt fractions are more or less typical of the eutrophic lake. The amino acid content of the sediments is like that of early eutrophic or oligotrophic lakes of the north central United States, but this is in part the result of more rapid inorganic sedimentation in Pyramid Lake than in the lakes of humic regions despite the probable greater rate of total sedimentation in the latter.

The humic acid constituents of Pyramid Lake sediments obtained by sulfuric acid hydrolysis also are in much smaller quantities than in lakes of Minnesota. Ultraviolet absorption spectra of fractions of the humic substances spearated by paper chromatography suggest the pyridines.

Comparisons of carbon and nitrogen content of various lake sediments in Minnesota with Pyramid Lake show that it ranks with the oligotrophic to early eutrophic types (table 7) in C and N content but is far above most eutrophic lakes in dissolved solids. For lakes of this type the term *apatotrophic* (Gr. *apatē*, deceit + *trephō*, feed) is proposed. The type is evidently characteristic of arid to semi-arid regions and owes its limnologic properties to a large influx of dissolved salts and a relatively high evaporation rate which inhibits development of all but a few adaptable plants and animals, thus keeping down the total productivity and C and N content. There is probably a fair balance between producers and consumers in such lakes although in Pyramid the organic content of the sediments suggests a deficit of consumers at the present time.

In discussing lakes of this type Hutchinson (1957) pointed out that "alkalitrophy" is present in lakes having at least 100 mgm per liter of Ca, a very hard water type of lake. The phytoplankton productivity in

such lakes is greatly restricted, even in areas rich in sources of P and combined N. Typical alkalitrophic lakes are those shallow Swedish high-calcium lakes, the Träske (Naumann, 1929, 1932), and are different in ratio of calcium to total alkalinity from Pyramid Lake. Sodium alkaline lakes such as Pyramid do not greatly restrict plankton growth (Hutchinson 1937, p. 108). As a result of these considerations it seems clear that the word apatotrophic as used here need not be confused with alkalitrophic types of lakes.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 298–315
FIGS. 1–14, SEPTEMBER, 1958

MASS MOVEMENTS IN PERMIAN ROCKS OF TRANS-PECOS TEXAS[1]

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ABSTRACT

Excellent examples of submarine toreva blocks, glide or slump folds, debris avalanches, and turbidity current deposits are exposed in the Permian rocks of the Guadalupe, Delaware, and Chinati Mountains. These structures are most common along the distal edge of reef talus that rims the Delaware and Marfa Basins.

INTRODUCTION

The Permian rocks of western Texas are probably not excelled for the abundance and variety of sedimentary structures related to deposition on a slope and to the adjustment of sediments to that slope. These structures are commonly developed within the sedimentary rocks at the margins of the large sedimentary and topographic basins that developed in western Texas, southeastern New Mexico, and northern Mexico during the Permian (fig. 1).

Various types of structure have been differentiated depending upon the time of movement with relation to sedimentation, the rate of movement, and the method and size of implacement and movement. Terminology applied is similar to that of subaerial mass movement.

Hydrology of the Delaware Basin and Guadalupe Mountains region has been described by King (1948) and more recently by Newell and others (1953). In both publications depth of water is postulated for various stages of reef development, based largely on height of the reef talus zone in front of the Goat Seep and Capitan reefs. Table 1, summarized from Newell and others. (1953, p. 190), indicates probable depth of water in the Delaware Basin immediately adjacent to the Guadalupe Mountains.

Unfortunately similar depth information is not available for outcrops in the Chinati Mountain, but primary sedimentary features in that locality are also considered to be formed in relatively deep water, well below the depth of wave action.

[1] Manuscript received March 29, 1958.

STRUCTURES RELATED TO DEPOSITION

Transported reef-talus blocks

Many large blocks and boulders of reef, reef-talus, and back-reef material occur in fine textured sediments of the basin margin adjacent to the Capitan and Goat Seep reefs of the Guadalupe Mountains. Small isolated erratic blocks of limestone are also present in fine siltstone and sandstone of the Ross Mine Formation in the Chinati Mountains (table 2).

Several blocks of reef material and reef-talus are exposed on the northeast wall of the mouth of McKittrick Canyon (fig. 2), immediately above the first road crossing, in the lower part of the Rader Limestone Member of the Bell Canyon Formation. These blocks illustrate well the distance to which large masses settled into soft underlying sediments (fig. 3). For example, one block sank approximately four feet into the soft underlying calcarenite. Underlying beds were deformed and a thick bed was pinched off beneath the block.

A particularly fine example of deformation caused by movement of a talus block is illustrated in the right side of figure 3. Beds are thinned over the top of the block and older ones are complexly folded outward from the block. Apparently as the block came to rest, after moving from left to right, it depressed the beds as it settled and deformed the beds in front of it into overturned and recumbent folds.

In addition to the blocks that are common at the base of the reef talus slope, isolated blocks of reef-rock occur in sandstone beds several miles from the scarp. Several such blocks of reef-derived material are clustered together at the base of the Rader

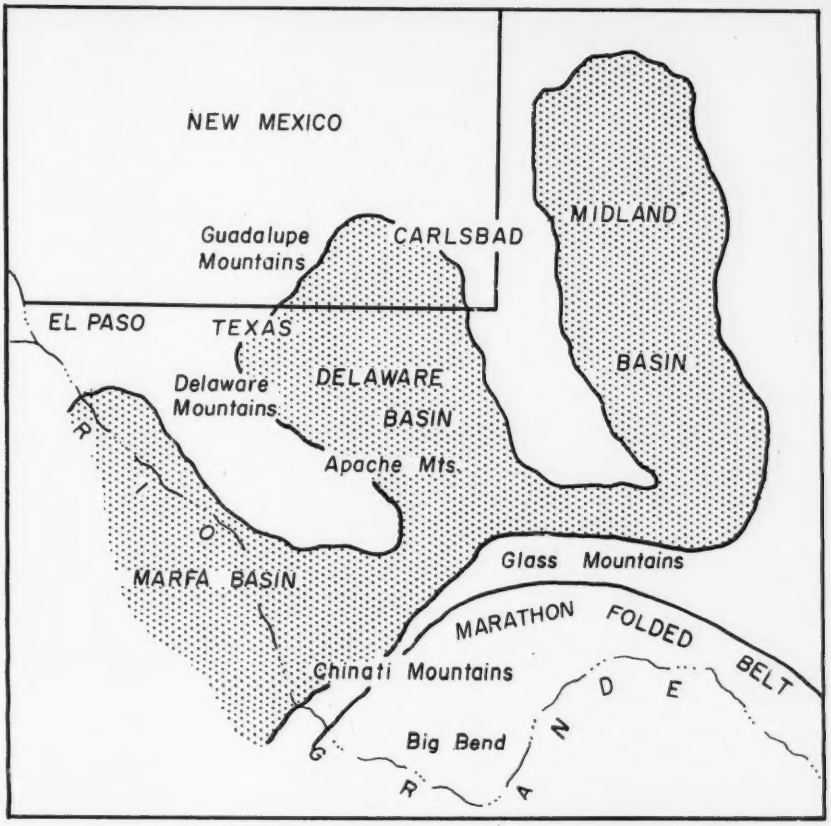


FIG. 1.—Index map of Trans-Pecos Texas and southeastern New Mexico. Stippled area includes deeper basins in contrast to surrounding reef and lagoonal areas.

Member, east of the main D Ranch road north of Bell Canyon. The blocks range from less than four inches to nearly six feet across. These blocks are probably contemporaneous with the basal conglomerate of the Rader Limestone found on Rader Ridge close to the reef scarp and basin margin.

In one of these blocks of limestone, laminae in adjacent fine-grained sandstone are depressed nearly two feet below the boulder, indicating the depth to which it sank in the water-laden sand before coming to rest. In an associated block, laminae have been depressed beneath the mass, but overlying laminae are essentially horizontal. The depth of settling of this block was probably only equal to the diameter of the mass.

Additional examples of well exposed displaced reef limestone blocks occur in the Pinery Limestone Member of the Bell Canyon Formation at the southeast base of El Capitan and in the narrow gorge of Pine Canyon (fig. 2). Rocks which enclose the Pine Canyon examples are thin to medium-bedded calcarenite and are complexly folded, presumably by impact of the block or by gliding that accompanied implacement.

TABLE 1.—*Depth of water in the Delaware Basin during Upper Permian deposition*

Formation or member	Approximate Depth of Water in Delaware Basin 2 miles from rim of basin
Castile Gypsum	1800
Bell Canyon Formation	
Lamar Member	1700
McCombs Member	1600
Rader Member	1600
Pinery Member	1500
Hegler Member	1100
Cherry Canyon Formation	
Manzanita Member	900
South Wells Member	1000
Getaway Member	
upper part	800
lower part	10–50
Brushy Canyon Formation	10–50
Bone Spring Formation	500

A large block of Lamar age (table 2) is exposed in three dimensions on the western face of a cliff on the south side of the mouth of McKittrick Canyon. Underlying strata have been laterally thickened by flow of sediment from below the settling rock. In the same general area, transported blocks of reef limestone lie in laminated sandstone below the McCombs Member in the stream bottom. Although the blocks are relatively large there has been little adjustment in the underlying strata. Blocks and boulders of similar reef limestone associated here are probably part of the same rock fall.

Isolated small boulders and cobbles of reef limestone are exposed in the reddish siltstone and laminated sandstone of the Ross Mine Formation west of Shafter in the southern Chinati Mountains. Distance of movement is not apparent from the small outcrops, but seems to be at least one mile judging from lithology of the overlying Capitanian rocks.

TABLE 2.—*Correlation table of Permian formations in the Guadalupe and Chinati Mountains, Texas*

SERIES		GUADALUPE MOUNTAINS			CHINATI MOUNTAINS	
OCHOAN		REEF AREA		BASIN AREA	CIBOLO CREEK	PINTO CANYON
GUADALUPIAN	CAPITANIAN	Capitan limestone	Bell Canyon formation	Lamar ls. McCombs ls. Rader ls. Pinery ls. Hegler ls.	Mina Grande limestone	Not exposed
	WORDIAN	Goat Seep limestone	Cherry Canyon formation	Manzanita ls. South Wells ls. Getaway ls.	Ross Mine formation	upper cherty limestone
				Brushy Canyon formation	not in contact	
LEONARDIAN		Victoria Peak limestone		Bone Spring limestone	Cibolo limestone	lower shale and siltstone
WOLFCAMPIAN		not exposed			Alta shale	not exposed

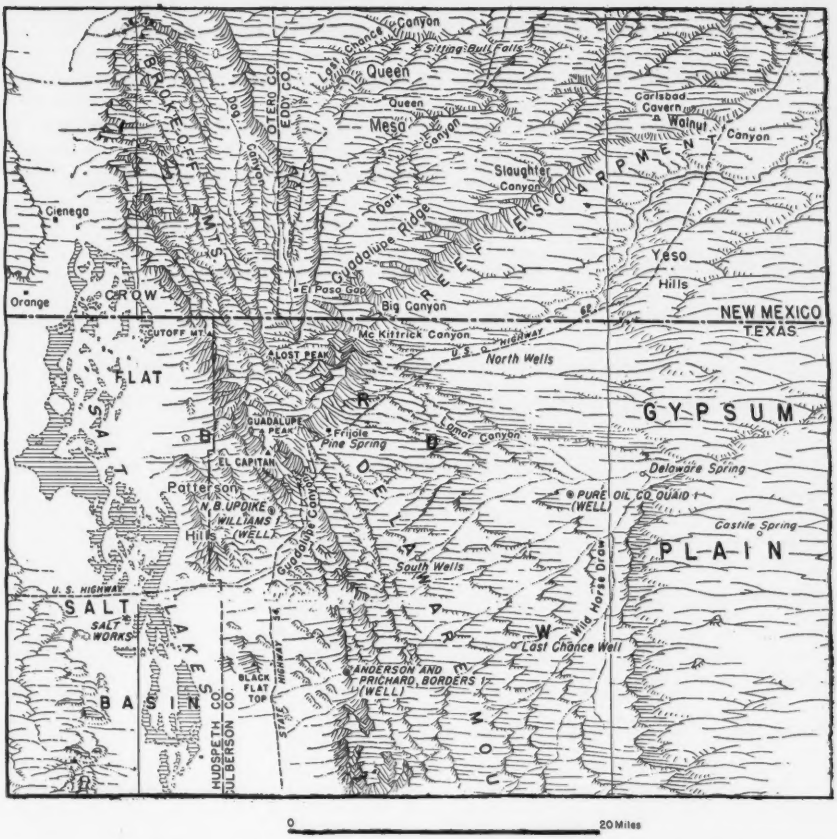


FIG. 2.—Physiographic index map of place names in Guadalupe Mountains and adjacent par of the Delaware Basin (adapted from King, 1948). B, Bone Springs; D, D Ranch headquarters R, Rader Ridge; W, West Chico Draw.

Debris Avalanches and Turbidity Current Deposits

Among the most impressive sedimentary features observed in the Permian rocks of Texas are the debris avalanches and associated turbidity current deposits that are exposed in the peripheral areas of the Delaware and Marfa Basins. Several such subaqueous mass movements are exposed in rocks along the base of the reef scarp in the Guadalope Mountains. Additional examples occur in outcrops of the southern Delaware Mountains and western Apache Mountains, at the southwestern margin of the Delaware basin, and in outcrops east of the main Chinati Mountains along the southeastern margin of the Marfa Basin. Although these deposits are submarine, classification used in the paper follows that of Sharp (1938), Thornbury (1954), and others because of the apparent rapid movement and heterogenous composition of the mass.

The Rader Member of the Bell Canyon Formation is composed largely of deposits of debris avalances and associated deposits (Newell, and others, 1953, p. 71–77). Three successive movements are seen in the member along Rader Ridge and southeastward

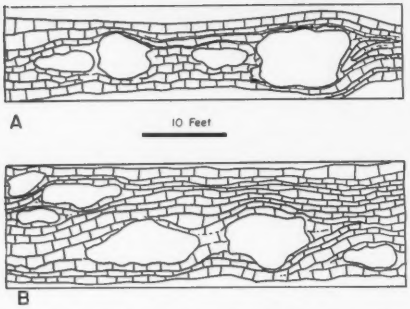


FIG. 3.—Erratic blocks of reef rock exposed in Radar Limestone at the mouth of McKittrick Canyon, Guadalupe Mountains. Blocks accumulated at the base of the reef talus in bedded calcarenite and moved from left to right.

in cuestas formed by the formation. The lowest of these is perhaps the most instructive and the best exposed example of debris avalanches seen during the present investigation. It is composed of a massive-appearing limestone breccia similar to the light-colored reef detritus of the Capitan Limestone, except that it extends basinward far from the toe of the reef talus. The mass forms a tongue within fine-grained basin sediments nearly 100 feet thick (figs. 4C, 5) and maintains this thickness for approximately one mile to the southeast of the talus slope where it terminates abruptly in a vertical face against bedded basin rocks one mile west of the Hegler ranch house fig. 4C). The upper part of the slide extends beyond this termination but is thinned to about 15 feet before terminating abruptly in massive sandstone and conglomerate (figs. 4A, C).

The angular or brecciated character of individual particles is obscure in most outcrops, but in places differential etching of the limestone surfaces shows the fragmental character of the rock. In most areas weathering obscures the texture of the rock if there is no contrast between detrital fragments and the matrix. Fragments of the breccia consist of very fine-grained calcarenite. Locally however, fragments composed of dark gray calcarenite are incorporated in the mass. Some fragments contain sycon sponges, large crinoid columnals and brachiopods. Sponges lie in parallel position within the blocks suggesting a direction of growth. There is little doubt that the sponge-bearing rock was derived from the Capitan reef and that it was transported with detrital fragments of talus material into the basin. Occasional fragments of limestone contain abundant *Mizzia*, a fossil blue-green algae, and encrusted fusulines. This association is characteristic of near-reef lagoonal rocks, and consequently, such blocks must have passed through a re-entrant in the reef, or collapsed with the reef crest at the initiation of the movement.

The massive limestone is continuous with the talus and lithologically almost identical with it. Silicification of fossils is nearly lacking in the reef and reef talus but is common in the upper part of the massive Rader breccia.

The abrupt termination of the breccia mass west of the Hegler ranch house affords a problem. Here, the abrupt change from light gray massive limestone breccia to thin-bedded sandstone takes place along a nearly vertical contact (figs 5, 6C). Superficially the contact resembles a fault, but limestone beds beneath and above it are not displaced. The relationship is either the result of deposition of sandstone against the massive limestone or of erosion of the sandstone followed by deposition of the breccia against the sandstone. The former possibility is unsupported by field evidence. Rader Limestone which intertongues with the breccia overlies the sandstone indicating that at least part of the breccia is younger than the adjoining sandstone. Other hypotheses for explaining the abrupt termination of the breccia have been seriously considered and abandoned. For example, it was suggested that the breccia mass may have been deposited in a previously existing depression, but the equally difficult explanation of the origin of the original depression complicates the picture. Another suggestion that deposition of the breccia may have been just equalled by deposition of the sandstone has been considered, but the boundary between the sandstone and breccia is sharp and there is no interfingering.

It is concluded that the avalanche itself excavated the depression as it fell into the soft sediments. Wherever the base of the breccia is exposed, fragments of the under-

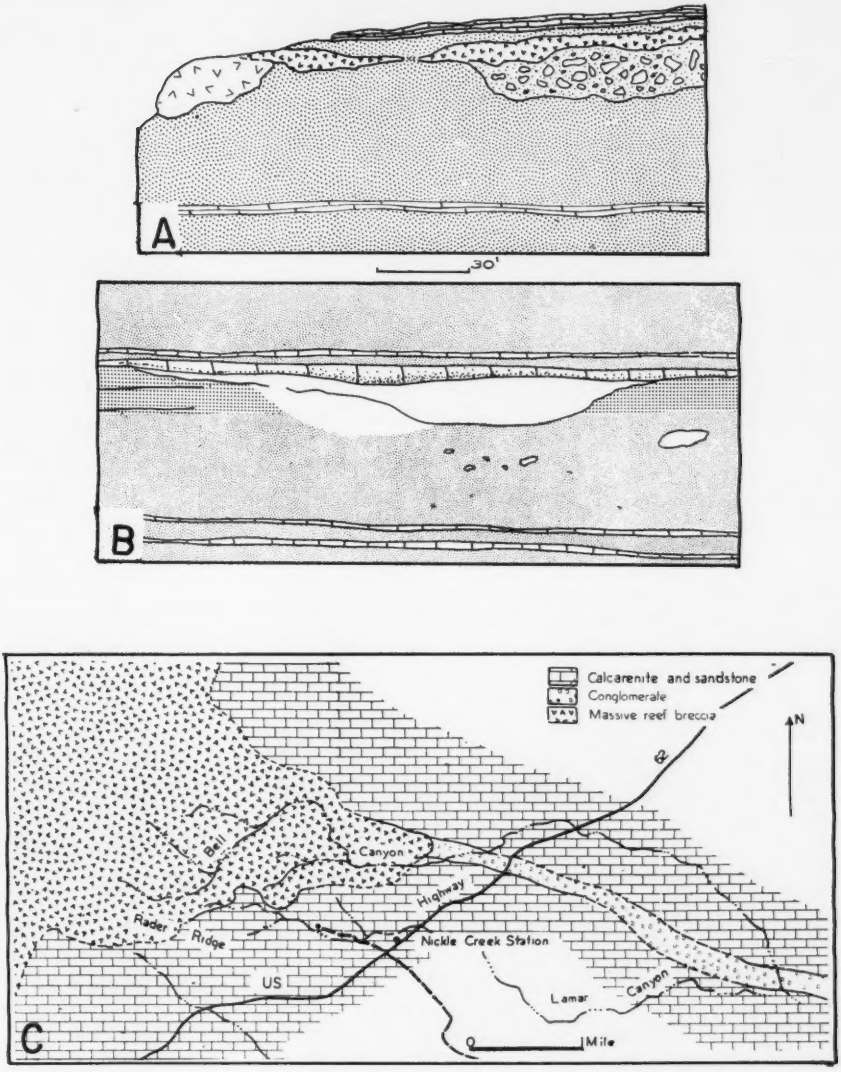


FIG. 4.—Rader submarine slide deposits (from Newell and others, 1953).

A. Rader slide breccias and overlying and underlying strata on north side of Bell Canyon near U. S. Highway 62.
B. Rader slide breccia on scarp east of "D" Ranch road near junction of Bell and Lamar Canyons.
C. Map of submarine debris fall and associated deposits of the Rader Member on Rader Ridge and along Bell Canyon.

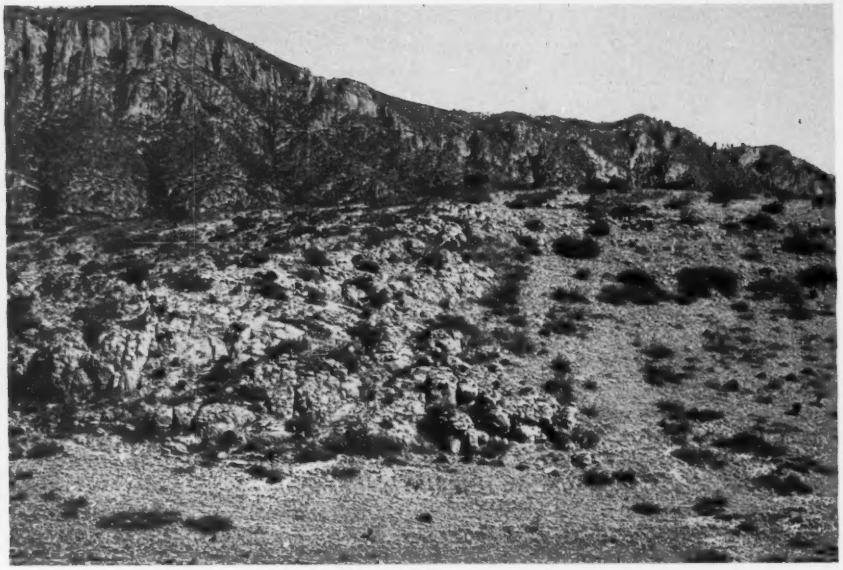


FIG. 5.—Rader massive limestone breccia as seen from southeast of Rader Ridge across Bell Canyon. Massive ledge in background is reef from which beds in foreground fell as submarine avalanche. (From Newell and others, 1953, by permission of W. H. Freeman and Company.)

ying sedimentary rocks are seen incorporated into the basal portion of the slide. The comparative lack of quartz sand within the mass, however, is difficult to understand since great quantities of sandstone must have been gouged out during implacement of the slide. The eroded fine-grained material apparently was swept into the basin as part of the finer component of a turbidity current. Quartz sandstone containing light-gray calcarenite boulders is the basinward extension of this slide.

Associated conglomerate and breccia extend the limits of the massive reef material (fig. 4C). Conglomerate of reef and reef-talus fragments in a quartz sandstone matrix lies beneath upper Rader calcarenite beds for several miles southeast and east of Rader Ridge. In part these conglomerates are contemporaneous with the massive limestone breccia but the others may be slightly younger (fig. 4A). The time lapse was very small but at least two separate rock falls are recorded in the lower Rader Member.

The sandy conglomerate which is considered directly contemporaneous with the massive slide is well exposed in a road cut along U. S. Highway 62, approximately 0.9 of a mile north of Nickel Creek Station. At this locality the conglomerate fills a channel several feet deep in bedded sandstone above the Pinery Member. Fossil fragments, including several species of brachiopods, fusulines, bryozoans, and corals, occur free of limestone matrix in the sandstone. These Permian rocks are associated with Quaternary caliche-cemented gravels and it is often difficult to distinguish the two. Permian fossils in the sandstone matrix indicates a Permian age for the channeled conglomerate.

The same conglomerate is exposed at several points east of the highway on the cuesta where it shows as a concentration of reef boulders and pebbles in a covered slope. At the intersection of Bell and Lamar Canyons (fig. 4C), near the D Ranch road, the conglomerate is represented by several large boulders of reef detritus ranging in diameter from the to six feet. Two periods of deposi-

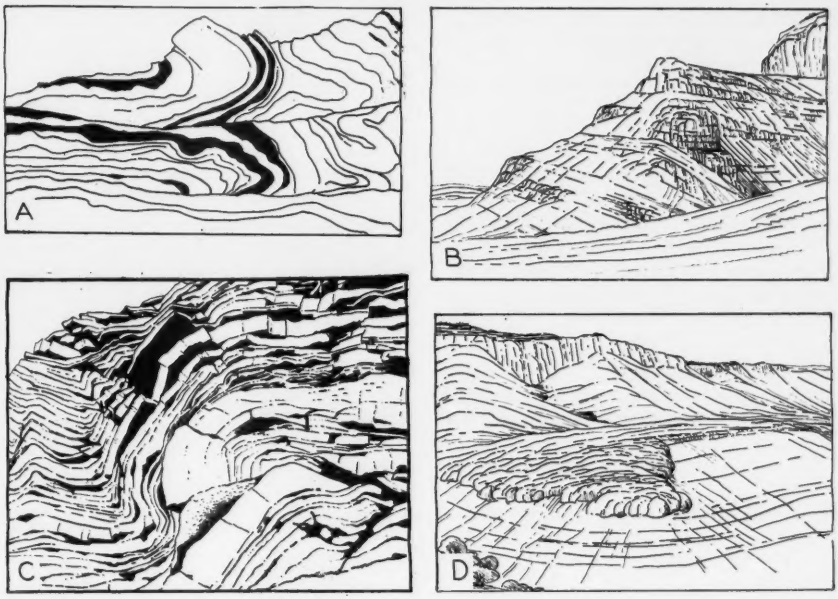


FIG. 6.—Primary mass movements

A. Slump folds in lower part of Wordian limestone in Pinto Canyon (fig. 10, locality F), with chert solid black, limestone white. Movement was from right to left, or from southeast to northwest toward center of Marfa Basin.

B. Submarine toreva block in Bone Springs Limestone, Shumard Canyon, along western side of Guadalupe Mountains, as viewed from southwest.

C. Slump fold in Lamar Limestone exposed on cuesta west of U. S. Highway 62, as viewed from southwest. Thick bed in center approximately 18 inches thick.

D. Field sketch of relationships of Rader submarine debris slide on Rader Ridge. Massive ledges are slide of light colored limestone breccia which ends in nearly vertical surface against bedded sandstone and limestone of Bell Canyon. Slide probably began at reef crest exposed near top of mountains in background.

tion of boulders are recorded here as at localities immediately west of U. S. Highway 62 (fig. 4B). The lower boulder bed is overlain by a thin sequence of poorly bedded quartz sandstones which in turn are overlain by an upper boulder bed. The lower slide is contemporaneous with the massive breccia of the lower Rader described earlier while the upper slide is equivalent to breccias and conglomerates that overlie the massive limestone on Rader Ridge.

It is evident from outcrops of lower Rader Limestone that turbidity currents can develop stronger currents in deep water (table 1) capable of eroding and transporting tremendous amounts of material down the sedimentary slope. Experimental work by Kuenen (1951, p. 29–30) has indicated turbibity currents are capable of eroding the floor over which they travel. Although his experiments were not conclusive to some students, they seem to be further substantiated by observations of these Permian examples where it is difficult to explain conglomerate-filled channels by any other means, particularly when the distal portion of the conglomerate is graded-bedded.

Basinward and southeast of Lamar and Cherry Canyons, thick accumulations of sandstone conglomerate characterize the Rader Member, particularly near the Pasotex pipeline and south to West Chico Draw (fig. 2). Near the latter locality the member is composed of thick contorted quartz sandstone and mudstone. The uppermost of four disturbed units is mainly thin-

bedded, black calcarenite, crumpled and contorted similar to upper Rader Limestone near Bell and Lamar Canyons. The upper unit is five to ten feet thick and fills a channel in the underlying beds. The lower three disturbed zones consist of buff to ash-gray sandstone which contains dark-gray limestone pebbles. Free fusulines are common and many are oriented northwesterly, parallel to the reef scarp. Green marly limestone which appears near the middle of the member to the north has been cut across in places by the upper disturbed zone.

These four beds of conglomerate may be approximately dated and related to mass movements close to the reef scarp by reference to the green limestone which serves as a datum. The upper crumpled limestone is equivalent to one observed above the lower massive Rader Limestone near Rader Ridge. Two of the lower disturbed conglomeritic beds are probably equivalent to disturbed rocks of lower Rader age developed on Rader Ridge. The third may be correlated with slides exposed to the southwest in Trew Canyon in the southern Delaware Mountains. Precise correlation is impossible because of the lack of adequate datum horizons.

Rader time was apparently a period of instability of the reef front and adjacent basin deposits in the Guadalupe Mountains, for at least three times of mass movement are recorded in the member on Rader Ridge and in Bell Canyon. The first of these during early Rader time involved collapse of a large segment of the Capitan Reef front. The second followed immediately afterward but was less intense. The third movement, or late Rader age, was not a submarine debris avalanche but a slumping of basin deposits after deposition, far down the slope from the talus toe. This upper slump will be described later in sections dealing with post-depositional sedimentary structures.

Deposits resulting from turbidity currents and submarine slump are also exposed on both sides of Trew Canyon (fig. 7) in the southern Delaware Mountains. There, two thick beds of coarse breccia and conglomerate show well the transition from a landslide deposit into that more characteristic of a turbidity current. Both these slides have been described earlier by the writer (in Newell and others, 1953, p. 69–72), but because of their stratigraphic importance they will be summarized here.

The lower of the two slides is of upper Cherry Canyon age and is roughly equivalent to the Manzanita Member of the Formation (table 2). The breccia forms a prominent cliff approximately half way up the side of Trew Canyon near its mouth, but descends toward the northwest until it crops out in the canyon bottom three miles north of the mouth. At the southern end of the exposure the breccia is approximately 30 feet thick but thins to approximately 14 feet before it disappears by faulting three miles to the north. Most of the fragments in the slide are subrounded to rounded and range from sand size up to more than four inches in diameter. The majority of the fragments are gray, fine-grained calcarenite, but light-gray and black fine-grained calcarenites are also common. The matrix contains much quartz sand and is greenish gray, a color characteristic of undolomitized Manzanita Limestone in the interior of the basin.

An important feature of the Manzanita Conglomerate is a gradual change from that showing little sorting in the southern outcrops near the source to well-graded beds farther away at the northern outcrops. Cobbles are concentrated near the bottom and fine-grained rocks near the top of the deposit, indicating strong graded bedding. Grading extends completely through the deposit without repetition or break in the northern outcrops, suggesting deposition of the complete unit as a single turbidity current deposit. Only in such a current would graded bedding be found throughout the unit in basinward extension while unsorted material accumulated close to the source.

The upper of the two breccia and conglomerate beds exposed in Trew Canyon (fig. 7) is roughly equivalent to the Rader Member slides described from the Guadalupe Mountains. The Trew Canyon deposit is approximately 40 feet thick where it crops out and appears rather uniformly that thick along the rim of Trew Canyon. The bed is composed of angular fragments of limestone, ranging from sand-size to blocks up to 15 feet across greatest dimension. Average fragments are approximately five

FIG. 7.—Trew Canyon in the southern Delaware Mountains, as viewed from the south. Cliff-forming cap rock and bench-forming unit in center of photograph are conglomeratic accumulations of Rader and Manzanita Members deposited as mobile submarine slides. Road in middle distance is along fault which drops rocks in foreground down relative to those behind. (From Newell and others, 1953, by permission of W. H. Freeman and Company.)

or six inches in diameter. These transported fragments are composed in part of light gray reef limestone, derived almost certainly from the reef front several miles to the south and southwest in the vicinity of the Apache Mountains. Other fragments are coarse to fine-grained, light-gray calcarenite that looks like reef talus. Most fragments, however, are lithologically most similar to calcarenite which elsewhere accumulated at the base of the reef talus slope. These rocks are now as much as ten miles from the nearest possible source at the basin margin. The matrix of fine-crystalline calcium carbonate forms probably no more than 5 to 25 percent of the rock, for the poor sorting produced close packing.

The base of the breccia mass produced little visible disturbance on the laminated and thin-bedded underlying sandstone and calcarenite. Channels and furrows are only minor shallow features. With the exception of small mounds (Newell and others, 1953, p. 74–76) developed on the upper surface, the top of the mass is parallel with the base.

Exposure of the Trew Canyon conglomerate masses is interrupted so that neither the source area nor the basinward extension of the beds is exposed. In both, however, there is a gradual decrease in size and increase in sorting when traced from south to north away from the basin margin.

Exposures of a coarse breccia bed were observed in the Chinati Mountains by early workers. Udden (1904) was the first to call attention to these rocks and termed the disturbed unit "the lower breccia bed." Baker (1927), Skinner (1940), and Rix (1953) followed Udden's usage. Udden described the rocks as consisting of light gray limestone in heavy ledges, often thoroughly brecciated. He pointed out that in some areas blocks six feet in diameter were noted. Udden's description was apparently taken from exposures near the junction of Cibolo Creek and

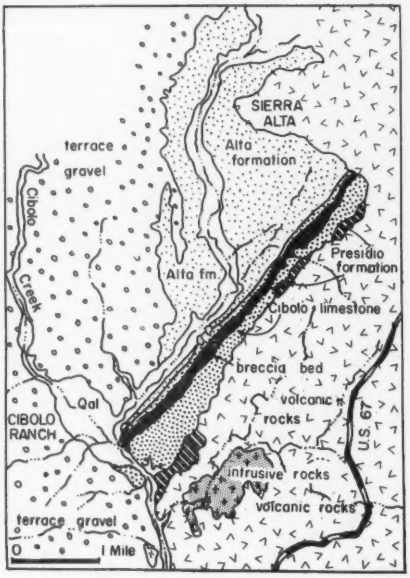


FIG. 8.—Geologic map of Cibolo Creek exposures of Permian rocks, Chinati Mountains, Texas (modified from Rix, 1953) showing distribution of breccia bed in Cibolo limestone.

Sierra Alta creek (fig. 8). To the northeast along the cuesta formed by the Cibolo Limestone additional exposures of the complete sequence occur where several small gullies cut into the face. Here coarse composition of the mass is readily apparent (fig. 9) and boulders up to one foot in diameter form the "fine" matrix between larger blocks. Isolated blocks up to 20 feet in diameter can be seen in nearly every exposure, and such blocks weather out to form a very coarse rubble on the partially covered slopes. The coarsest block seen is exposed near the eastern end of the outcrop mass, due south of Sierra Alta (fig. 8), where a single block approximately 900 feet long and 40 feet thick is surrounded on the cuesta face by coarse breccia.

Most of the fragments within the mass are composed of light colored calcarenite and minute breccia characteristic of the reef talus zone. Fusulines and algal masses are common in some blocks suggesting origin at edge of a reef escarpment. Sponges, brachiopods, and crinoidal debris occur in other blocks in what would be called a reef assemblage. Not all the blocks are light colored; a few are composed of darker colored basin-type calcarenites. Most fragments are relatively well rounded but some are distinctly angular. Matrix between th· finer fraction of the mass is formed by granular calcite and quartz sand. The matrix is characteristically light yellow-gray to brown-gray. It is relatively easily weathered and leaves the boulders in strong relief, stained light yellow gray.

The disturbed zone is approximately 130 feet thick throughout its limited outcrop, being slightly thicker toward the eastern end of the belt (fig. 8). Graded bedding is limited to the upper five or ten feet where it is well developed. Apparently the mass is too coarse grained in the lower part for grading bedding to show. It is also possible that the breccia as exposed is not far removed from the source, for overlying rocks in Cibolo Narrows are in well-defined reef facies. Direction of movement is not known but is assumed to have been from southeast to northwest from the basin margin. Sedimentary structures below and above the breccia bed indicate the sedimentary slope was toward the northwest into the Marfa Basin.

It is possible that disturbances in Pinto Canyon which produced submarine toreva blocks and slump folds are contemporary with the breccia in Cibolo Creek and that both may be roughly equivalent to similar features developed in the Bone Spring Limestone of the Guadalupe Mountains.

STRUCTURES DEVELOPED FOLLOWING OR PENECONTEMPORANEOUS WITH SEDIMENTATION

Submarine toreva blocks

Large tilted blocks and wedges of limestone and black shale and sandstone are exposed in Leonardian rocks of the Guadalupe and Chinati Mountains. Smaller but similar tilted blocks are exposed in the Lamar Limestone Member of the Bell Canyon Formation along the front of the Guadalupe Mountains near the Texas-New Mexico state line. Although early workers in the Guadalupe Mountains attributed the irregular attitudes of some wedges to gigantic cross-bedding (Lloyd,

FIG. 9.—Coarse-textured upper portion of slide breccia exposed in the lower part of Cibolo limestone along Sierra Alta Creek, eastern Chinati Mountains, Presidio County, Texas. Bedded limestone at top grades into fine sponge spicules from coarse debris below.

1929, p. 657) or to irregular local unconformities which resulted from tectonic activity (Darton and Reeside, 1926, p. 423; King, 1948, p. 15), recent work by Newell (1953 p. 86–87) indicates that these wedges are slumped blocks. Because of their probable mode of origin and apparent backward rotation as slumping occurred, these features are considered to be toreva blocks that moved during and immediately after deposition in a subaqueous environment. Examples in Leonardian rocks (fig. 6B) in the Guadalupe Mountains have recently been described by Newell (1953, p. 86–89).

Large wedges of sedimentary rocks, similar to those exposed along the western scarp of the Guadalupe Mountains, are exposed in the Leonardian black shale and sandstone sequence in Pinto Canyon in the northwestern portion of the Chinati Mountains. One excellent example is exposed immediately above the next to the last crossing of Pinto Creek, approximately three-quarters of a mile upstream from the Perez Ranch house (fig. 10, locality D). The base of the block is well exposed in the bottom of the creek (fig. 11). The base is disconformable, with approximately 10 to 15 degrees divergence between the block and the underlying sedimentary rocks. A small fault cuts the slump surface immediately before the base of the block is buried beneath younger terrace gravel. A block of sediments approximately 150 feet thick is involved in the movement for the folded top of the block is exposed approximately a thousand feet upstream from its base (fig. 10, locality C). The sharp break with little brecciation and folding at the base seems characteristic of this type of slump structure.

Other large wedges are exposed elsewhere in Pinto Canyon. A large block, approximately 200 feet thick, is exposed on the south wall of Pinto Canyon, due south of the Perez ranch house (fig. 10, locality G). Several smaller blocks are also exposed in the canyon bottom and walls in the upper part of the organic siltstone and sandstone sequence. One example is particularly well exposed in the creek bed approximately two

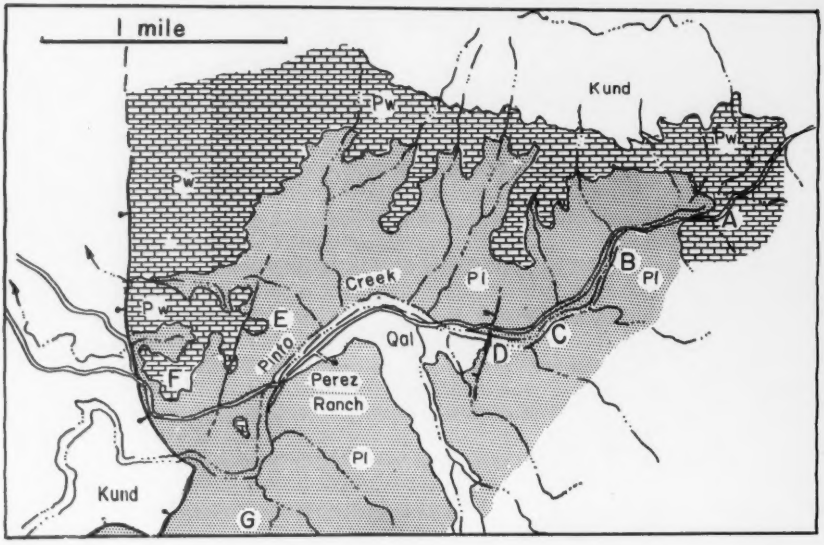


FIG. 10.—Geologic map of Pinto Canyon, northwestern Chinati Mountains, Presidio County, Texas. Stippled pattern is outcrop area of Leonardian siltstone and sandstone sequence. Limestone pattern is outcrop area of lower Wordian limestone and chert sequence.

miles above the ranch, and demonstrates the same smoothly sheared base characteristic of the larger blocks.

A single example of a similar slumped mass crops out farther upstream in Pinto Canyon, in the lower part of the Wordian Limestone sequence (locality A), where chertified limestone and fossiliferous black calcarenite have slid as a single twenty-foot block.

With the possible exception of the examples in the Bone Spring Limestone of the Guadalupe Mountains, the slope necessary to induce toreva block movement cannot be determined with certainty. Newell (1953, p. 86–89) concluded that structures developed in the Bone Spring Limestone are the result of movement off the flank of the Bone Spring flexure as it was rising in mid-Leonardian time. He states that the regional dip of the flexure is as much as 15 degrees and that the dip gradually decreases basinward to only one or two degrees. The maximum angle of dip off the flexure is comparable to figures obtained by Kuenen (1935, p. 73) as the maximum angle of slope upon which clay and ooze may accumulate in relatively thick layers. In all the above examples, the sediments are highly organic-rich and fine textured, and when water laden may have slumped at angles considerably less than ten degrees. Unlike the Guadalupe Mountain area, Chinati Mountain examples apparently developed far from the basin margin, suggesting movement on very low slopes, possibly of the order of two or three degrees.

Slump Folds

Slump folds are primary structures developed in unconsolidated or partially consolidated sediments as a result of slump, creep, or other lateral movement of surface or near surface sediments. They include both those folds within the slump mass and the folds in adjacent strata caused by impact or drag of the mass. Folds considered to be of this type are abundantly developed in the Guadalupe and Chinati Mountains. Slump folds are abundantly developed in the Bone Spring Limestone, in limestone of the Lamar and Rader Members of the Bell Canyon Formation, and in limestone of the Getaway Member of the Cherry Canyon

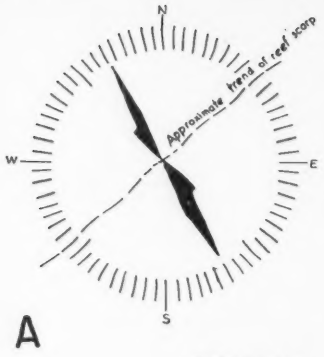
FIG. 11.—Base of submarine toreva block exposed in Pinto Canyon (fig. 10, locality D), in organic rich siltstone and sandstone of Leonardian age. Scale shown by hammer on lowest ledge of block. A small fault, later intruded by dike, cuts western margin of exposed toreva block.

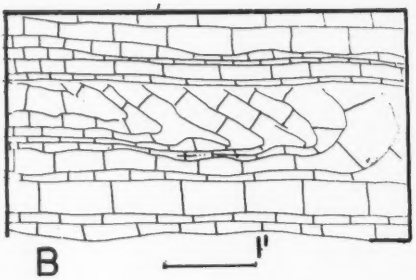
Formation along the eastern flank of the Guadalupe Mountains and in the northern Delaware Mountains at the margin of the Delaware Basin. Excellent examples of slump folds are exposed in Pinto Canyon in both the upper and lower units of the Permian and in the Alta shale along the eastern side of the Chinati Mountains near Sierra Alta and Cibolo Creek.

Summary observations and illustrations of some of the Guadalupe Mountains occurrences have been published elsewhere (Newell and others, 1953). In the same volume Newell (1953, p. 86–89) has described the folds developed in the Bone Spring Limestone of the northern Delaware Mountains. Little can be added to his description of the structures and to their illustration. Additional information can be summarized, however, for slump folds developed higher in the section.

Intricate deformation of fine-grained calcarenites of the Lamar Member is developed for several miles along the outcrop belt of the member southeast of McKittrick and Big Canyons (fig. 2). One of the largest folds observed is exposed along the Watkin's ranch road, near the Texas and New Mexico state line at the mouth of Big Canyon, where a thickness of at least 60 feet of beds is disturbed. The top of the folded mass has been removed by erosion, but the internal structure and basal relations are preserved. The bedding is well preserved in complex recumbent folds and is not appreciably modified by pinching or swelling. Deformed beds overlie undeformed strata along a flat sedimentary thrust surface. The gliding surface of the overlying mass corresponds to a bedding plane but truncates the underlying limestone at a low angle, varying from 15 to 20 degrees.

Contact drag folds and slickensides are lacking along the glide-surface. Freedom from angular pebbles at the contact indicates that the sediments were not lithified at the time of deformation; it is difficult to understand the general lack of friction structures at the slump surface. The underlying mass has not been altered but the overlying mass is folded with individual





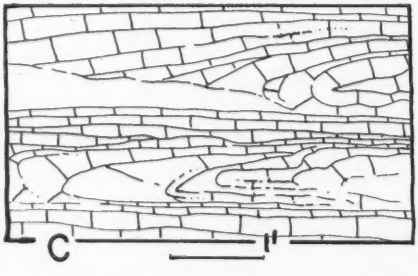


FIG. 12.—Lamar slump folds. (From Newell and others, 1953.)

A. Regional alignment of primary folds in the Lamar Member as exposed southeast of U. S. Highway 62.
B, C. Recumbent slump folds exposed southeast of U. S. Highway 62, north of "D" Ranch road and Bell Canyon.

folds overturned to the northeast indicating movement in that direction. The axial trend of most of the folds in the outcrop is approximately N 30° W, a direction that is similar to trends of other folds in the Lamar Member farther out in the basin to the southeast (fig. 12A).

Another excellently exposed slump fold occurs in the upper part of the Lamar Member on the first promontory west of U. S. Highway 62, north of the road to the McCombs and Pratt ranches and McKittrick Canyon (fig. 4C). The base of the deformed sequence is not exposed, but undisturbed laminated sandstone is visible a few feet below the crumpled limestone. Mildly deformed strata crop out on the crest of the hill. Deformation gradually increases in intensity upward (fig. 6C) indicating that only the lower portion of the moving mass was folded by drag along the slump surface. The underlying organic-rich sandstone evidently formed the slumping surface.

Differences in reaction to stress of unlike beds are apparent. A relatively thick bed at the center of the outcrop was fractured and thrust approximately three feet. A prominent drag fold is developed in this bed, but overlying and underlying rocks are more intensely folded. The thick fractured bed is composed of coarse-grained fossiliferous calcarenite. Enclosing thin beds are finer textured and only sparingly fossiliferous. Evidently the black fine-grained beds were relatively plastic during deformation. Angular folds occur in several places but none was cut by contemporaneous fractures.

Other small folds of Lamar Limestone are exposed in the member southeast of the highway (fig. 12B, C). Each of these folds shows overturning or drag indicating movement toward the northeast. The remarkable alignment of these small folds is shown in fig. 12A, a compilation of trends of fold axes along the cuesta of the Lamar limestone southeast of the highway. Almost all the folds in the Lamar Member trend nearly normal to the reef front. Presumably the folds were formed by movement downslope. These folds may have resulted from northeast tilting of the basin floor, perhaps by differential subsidence of the area northeast of the Huapache monocline which cuts through the Guadalupe Mountains in the

vicinity of Big Canyon and Slaughter Canyon with a trend approximately normal to that of the reef trend (see King, 1948).

Disturbed upper Rader Limestone beds crop out in Bell Canyon a short distance below its confluence with Lamar Canyon. The disturbed rocks consist of complexly folded slumped calcarenite preserved in a channel cut in underlying sandstone (fig. 4B). Several feet of laminated sandstone occur between this mass and the underlying boulders thought to be the same age as the submarine debris slide of the Lower Rader on Rader Ridge. The slumped limestone represents the upper of three slumped deposits of the Rader Member.

Limestone involved in the Upper Rader slump mass seems to have been plastic at the time of movement, since it is complexly folded with little fracturing. The limestones piled up in the channel are probably part of a larger slide mass. Since reef blocks are lacking in the slump it must have originated in the fans below the talus zone at the base of the reef scarp.

The effect of slumping in the upper Rader Limestone can be seen in the basin for a considerable distance; the calcarenites are thrown into folds as far as the member crops out, some 28 miles from the reef scarp.

This last slump of the Rader Member was evidently a surface feature on the sea floor. The base of the disturbed zone is nearly smooth, but locally channeled, and the top is planed by a very persistent conglomerate that exhibits well-defined graded bedding. Large pebbles are concentrated at the base and fine-grained calcarenite at the top (fig. 4B). This bed of conglomerate extends over many tens of square miles in the outcrop area. Such a bed might be considered the result of a turbidity current set up by the slump. It exhibits most of the characters commonly ascribed to such deposits (Kuenen, 1951, Kuenen and Migliorini 1950).

Deformed strata are exposed in the Getaway Limestone Member of the Cherry Canyon Formation near Pine Spring Camp and in most exposures of the Upper Getaway Limestone along both sides of Guadalupe Canyon (fig. 2). Near Pine Spring Camp, a thin-bedded fine-grained calcarenite is deformed in small recumbent folds with approximately 12 feet of structural relief. Individual folds die out vertically

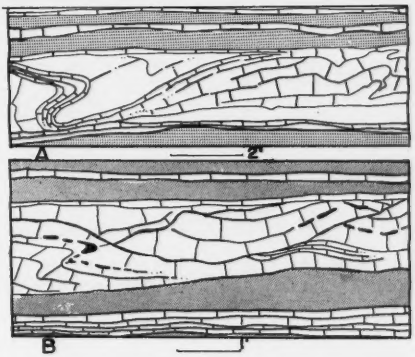


FIG. 13.—Primary slump structures in Pinery Member of Bell Canyon, exposed on Rader Ridge near Bell Canyon. (From Newell and others, 1953).

within a few feet. They were probably formed in plastic calcareous muds beneath a light cover of sediment. There is no evidence that the folds reached the surface and alignment of fold axes is poor. Similar poor alignment is developed in numerous folds within fine-grained black dolomite of the upper Getaway Limestone on the east side of Guadalupe Canyon.

Small folds in lenticular limestone beds of the Pinery Member of the Bell Canyon Formation are also considered to have originated as slump features (fig. 13). Their development is most consistent near the base of the reef talus on Rader Ridge, where nearly every limestone bed is mildly deformed.

There are several points in favor of slumping as the cause of the Bell Canyon and Cherry Canyon folds.

1.—In the first place, large masses or sheets of sediment are involved. As pointed out earlier, over 60 feet of strata are involved in the Lamar folds near Big Canyon. It seems inconceivable that strata of such thickness could have been deformed in place by differential loading and intrastratal flow.

2.—Perhaps the strongest argument in favor of sumping or creep as the cause of these folds is the nearly perfect alignment of fold axes in the Lamar Member. Such an alignment, it is believed, demands that these folds were formed as integral parts of one mass movement.

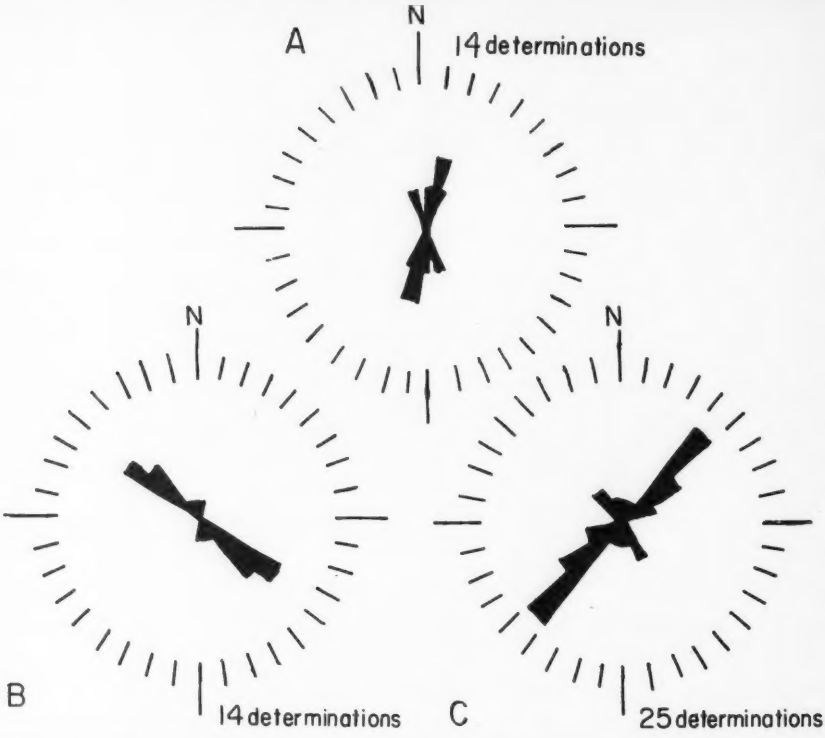


FIG. 14.—Alignment diagrams of fold axes exposed in Pinto Canyon, northwest of the Chinat Mountains, Texas. A. Alignment of fold axes in lower part of Wordian Limestone, upper Pinto Canyon, fig. 10, locality A; B. Alignment of fold axes in lower part of Wordian Limestone unit, lower Pinto Canyon, locality F; C. Alignment of fold axes in Lower Wordian Limestone unit, lower Pinto Canyon, locality E.

3.—The Rader Limestone is disturbed over an area extending for more than 25 miles along the outcrop band. Such widespread disturbance strongly suggests slumping as the cause. Unfortunately, measurements of the axes of Rader folds were not made.

4.—The graded conglomerate at the top of the Rader Limestone is adequate evidence that the movement which crumpled these rocks occurred at the surface of the sea floor.

Slump folds were observed in organic-rich sandstone and shale of Leonardian age and in siliceous calcarenite of Wordian age in Pinto Canyon of the northeastern Chinati Mountains, and along Sierra Alta Creek, east of the mountains.

Leonardian examples of slump folds are particularly well exposed in the upper reaches of Pinto Canyon, approximately two miles upstream from the Perez ranch house (fig. 10, locality B). Here sandstone and siltstone, with minor beds of calcarenite, are complexly folded and broken. Shale and siltstone beds behave as incompetent units, swelling and pinching but very rarely fracturing. Interbedded calcareous sandstone and silty calcarenites are competent and fracture relatively easily. Strike of fold axes vary somewhat (fig. 14A) between N 45 E to N.-S, but cluster between N 10 to 25 E. Movement of the mass was evidently toward the northwest into the center of the Marfa

Basin. Small sedimentary thrusts developed in the northeastern part of the outcrop have the same trend.

Siliceous cherty limestone beds which form the upper Permian rocks of Pinto Canyon display slump folds in most localities. Excellent examples are exposed at localities E and F, on figure 10. Trends of fold axes vary in each locality (fig. 14 B, C) but show a preference for northerly and northwesterly movement. Slumping produced overturning and rolling of beds, particularly at locality E, rolled limestone masses in the slumps appear as boulders or blocks. Bedding is accentuated in some masses by laminated and interbedded black chert, producing a charred or clinkered appearance when viewed from a distance.

Both the Leonardian and Wordian Pinto Canyon slump folds are associated with submarine toreva blocks. In both structures, movement was toward the north and northwest, presumably into the Marfa Basin from its southeastern margin.

The precise angle of gliding or slumping in the various areas can not be determined because of subsequent diastrophism. However, stratigraphic evidence of the thickness of the Castile Formation in the Delaware Basin indicates that the angle at which gliding took place in the Guadalupe Mountain occurrences was very low. The base of the Castile was probably nearly a horizontal surface in later Permian time. Other described slump folds also apparently moved in areas of low sedimentary dip (Hahn, 1913; Miller, 1922; Heim, 1908; Bradley, 1931). The Lamar slump folds probably occurred on a slope of less than two degrees and other slumps probably occurred on a similar angle.

ACKNOWLEDGMENTS

This paper is largely an outgrowth of a joint investigation of Permian reefs undertaken by graduate students of Columbia University under the direction of N. D. Newell. Among those whose helpful discussions and criticisms greatly aided the investigation are included: A. G. Fischer, A. J. Whiteman, J. S. Bradley, D. B. Blake, Ellis Yochelson, H. N. Fisk, W. H. Bucher, Marshall Kay, F. G. Stehli, J. P. D. Hull Jr., W. W. Wiles, and H. J. Bissell, in addition to N. D. Newell.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 316–320
FIGS. 1–7, SEPTEMBER, 1958

SEDIMENTARY "BOUDINAGE" STRUCTURES IN THE UPPER DEVONIAN IRETON FORMATION OF ALBERTA[1]

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ABSTRACT

The inter-reef Ireton Formation of the Upper Devonian Woodbend Group of Alberta is essentially a calcareous shale with variable amounts of inter-bedded nodular limestones. Structures formed in limestone of the Ireton Formation vary from thin relatively undeformed beds to pinching and swelling ones, and finally to completely isolated lenses or nodules that are variously oriented. The nodules consist of finely granular lime (fine sand to very fine silt size) with minor quartz in the same grain size range as the calcite, and variable amounts of clay. All nodules are surrounded by finer grained less limey rock and display fine, peripheral, calcite-filled cracks. These structures are thought to be "sedimentary boudinage" formed by compaction. Thin limey beds sandwiched between more clayey and hence more plastic beds, are pulled apart by the laterally moving plastic beds when subjected to unconfined compaction near the depositional interface. Continued compaction under hydrostatic conditions further modifies these structures.

INTRODUCTION

Bedding is typically disrupted in the calcareous shales and argillaceous limestones of the Upper Devonian Ireton Formation of Alberta by pinching and swelling and the development of isolated nodules. This type of structure is believed to result from pulling apart the more brittle limey strata by relatively plastic shales during compaction and is similar to the boudinage or sausage structures of metamorphic rocks.

Newell and others (1953, p. 80–81) noted beds irregularly and abruptly thickening and thinning and even wedging out completely to form lenses a few feet long, in the fine grained basin limestones of the Permian in the Gaudalupe Mountains. They attribute this to differential loading and intrastratal flow. Natland and Kuenen (1951) described "pull-apart" structures in both sand and clayey strata of the Ventura Basin in California. Beds have been torn apart into slabs which resemble "boudinage" but have less regularity, not much thinning and less indication of flow in underlying material. These structures appear to have resulted from contemporaneous sliding on a gentle slope or from drag exerted by turbidity currents. Hadding (1931) noted that even horizontal strata can slide if they are differentially loaded, as in a growing delta where overburden presses aside the plastic substrata. He described mostly highly complex folds resulting from intrastratal flowage. During this process strata were thinned in some places and thickened in others and may have become completely separated. Osmond (1956) noted that dolomite may contain irregular lenticular mottles of anhydrite, and he thinks that these may represent boudinage remnants of thin laminae separated by compaction forces.

Lombard (1956, p. 347) quotes Bellière on structures similar to those in the Ireton, from the Famennian of Belgium but they are attributed to an original lack of continuity in the bedding with stretching and thickening in a fluid paste which consolidated later.

[1] Manuscript received April 4, 1958.

GENERAL STRATIGRAPHY

Part of the Ireton and the underlying Duvernay Formation are an off-reef shaley equivalent of the Leduc reef carbonate formation. The lime contained in the middle and lower parts of the Ireton occurs as very fine clastic grains derived at least in part from the statigraphically equivalent reefs. The lime in the upper Ireton is autochthonous and derived mostly from sessile benthonic organisms.

The Ireton consists of a lower unit of about 150 ft of light grey-brown argillaceous limestone; a middle unit of about 250 ft of non-fossiliferous calcareous green-grey shale with only minor thin beds of

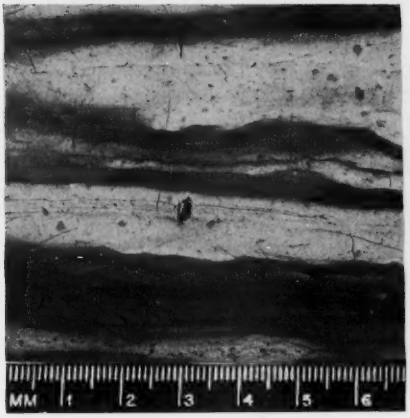

FIG. 1.—Pinching and swelling beds of argillaceous limestone, polished section, middle Ireton, Imperial Kingman (12-9-49-19W4), 4440 ft.

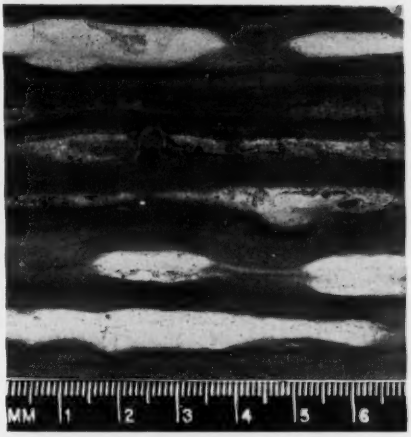

FIG. 2.—Limestone nodules in calcareous shale (sedimentary "boudinage"), polished section, middle Ireton, Imperial Kingman (12-9-49-19W4), 4445 ft.

limestone; and an upper unit of about 200 ft of thinly bedded very fossiliferous limestones and green-grey shales. This formation is known only in the subsurface, and the specimens studied were all obtained from drill cores. These were selected over a stratigraphic interval of 500–600 feet from several wells in the general Edmonton area.

DESCRIPTION OF IRETON STRUCTURES

Structures in these sediments stand out by virtue of compositional and textural differences. All of the limey layers and nodules are lighter colored and coarser grained than the surrounding sediments. The more calcareous thin beds or laminations in the Ireton are characteristically nodular or wavy. Limestone masses vary from thin relatively undeformed to slightly pinching and swelling (fig. 1) beds to completely isolated lenses lying dominantly parallel to the bedding (fig. 2) and irregularly oriented lenses or "mottles" (fig. 3). The coarser grained less argillaceous nodules of the lower Ireton are less elongated than those of the middle Ireton (figs. 2 and 3). All of these structures are thought to be similar in origin and differ only in degree of development.

Beds that extend completely across a core may pinch out close beyond because other similar layers terminate within the core boundaries (fig. 2). The lenses are irregular

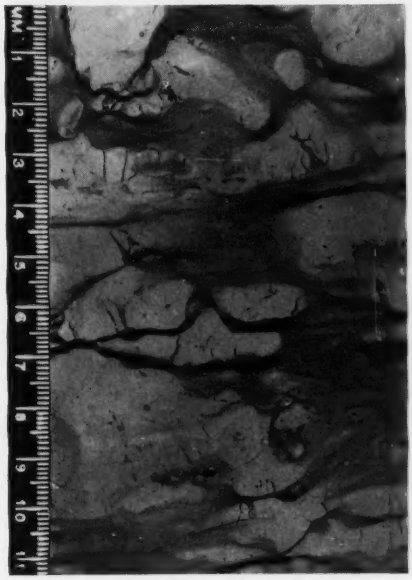

FIG. 3.—"Boudinage" with well developed tension cracks. Typical structure of lower Ireton. These nodules formed of purer limestone than those in fig. 3, polished section, lower Ireton, Imperial Labyrinth Lake (15-14-48-23W4), 5523 ft.

FIG. 4.—Three dimensional view of limestone "boudinage" in less limey matrix, polished block, Lower Ireton, Gulf Porter (16-31-41-18W4), 5135 ft.

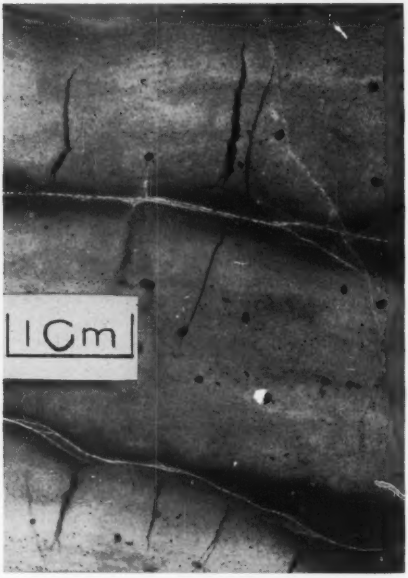

FIG. 5.—Lenses with well developed tension cracks, polished section, lower Ireton, Imperial-Canadian Superior Buck Creek (14-29-48-6W5), 8655 ft.

to ovoidal in plan (fig. 4) and generally range from 5 to 20 mm thick. Long axes generally are horizontal except where the lenses are draped over one another. Even some fine laminations show a tendency to develop this type of structure.

Fine wedge-shaped cracks which are filled with clear coarsely crystalline calcite

FIG. 6.—Lenses of argillaceous limestone in calcareous shale; lenses contain much greater proportion of lime grains which are coarser. Texture is indicated by fine pits representing removed lime grains. Etched section, middle Ireton; cut from same sample as fig. 1.

and narrow inwards from the borders are characteristic of the lenses (figs. 3, 4, 5). They are best developed and are most abundant in the purer, coarser grained limestones of the lower Ireton unit. In horizontal section they are irregularly distributed. The cracks are generally less than 1 cm in vertical length and less than 1 mm in width.

TEXTURE

The Ireton limestone beds and lenses possess a distinctly granular texture. Grains consist of lime with traces of quartz and range from fine sand to very fine silt size. They are angular to subrounded, and lie in a matrix of clay and very fine grained calcite. Recrystallization has been slight and is confined to the matrix. The nodules of the middle unit generally are finer grained and more argillaceous than those of the lower Ireton. The calcareous shale or very argillaceous limestone is invariably very much finer grained (clay sized) than the included lenses, and contains a much higher proportion of clay minerals. This contrast in grain size can readily be seen on the etched surface of figure 6.

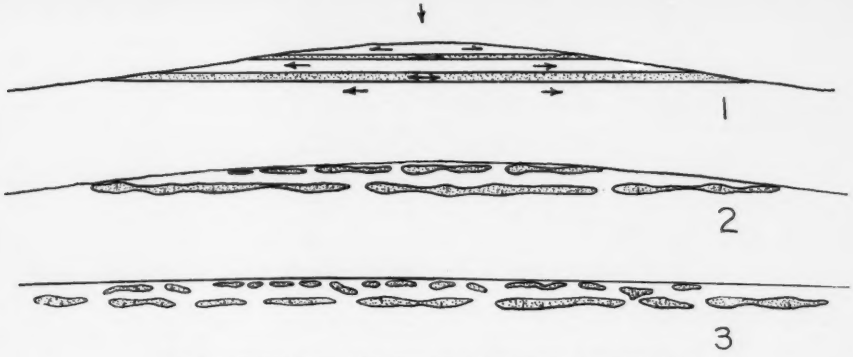

FIG. 7.—Stages in the development of sedimentary "boudinage."

ORIGIN OF STRUCTURES

The following sequence of events is suggested to account for the structures present in the Ireton Formation.

Lime clastics of fine grain size were deposited with varying proportions of clay in thin beds on a gently undulating bottom. As these fine muds, unconfined laterally, were subjected to compaction, they spread plastically towards the slightly lower parts of the sea bottom or they might have flowed laterally as a result of density variations in the sediment (fig. 7). The more plastic clayey beds dragged the surface of and created tensional stress in coarser grained, limey beds. As the unconfined mass spread laterally, the limey beds began to thin at points of weakness until eventually they ruptured to form separate lenses. As the sediment was more deeply buried and subjected to hydrostatic pressure, the lenses became more rounded as shale was squeezed around them. If lenses were abundant and closely spaced, they interfered with each other during compaction and were draped over and squeezed between each other. The final result in some beds was limey lenses scattered randomly through shale.

Such a process seems plausible for several reasons. Skempton (1945, 1953) showed clearly that the plasticity of a granular sediment increases with increasing clay content. The lenses in the Ireton are more limey, coarser grained and less plastic than the surrounding more clayey sediment. Ramberg (1955) noted that boudinage structures may develop in a competent layer sandwiched between incompetent ones when acted upon by compressive stresses perpendicular to the layering. Plastic flowage then takes place in the incompetent layers which creates tensile stress in the competent layer as long as there is some friction at the interface. The stresses formed in the limey beds or competent layers will depend on many variables, such as magnitude of compression rate of flow of plastic layers, viscosity of layers, amount of slippage between layers, area of contact, and thickness of layers. The upper limit of stress will be the ultimate tensile strength of the limey layer. Variations in these parameters may explain the somewhat different shapes of the structures of the coarser grained lower Ireton limestones and the finer grained more argillaceous limestones of the middle Ireton. Ramberg produced these conditions experimentally and formed structures very similar to those occurring in the Ireton cores.

Every gradation is found between undeformed beds and irregular lenses scattered randomly in the shale. The calcite-filled cracks suggest that the limey beds have been tensionally elongated. The finely clastic nature of the lenses indicates that this material probably was originally deposited in more or less continuous thin beds or laminations.

The hypothesis presented depends on the lateral spreading of the soft sediment, and therefore requires the assumption of a slightly irregular depositional surface or slight lateral variations in the density of the mud. The relief would only have to be very

slight to provide for lateral motion.

The Ireton nodules do not appear to have resulted from any other penecontemporaneous process. Some of them resemble concretions superficially, but their texture is sedimentary and they do not display concentric or radial structure. The literature on concretions contains little or no information on textures and it is conceivable that some nodules said to be concretions are in reality sedimentary "boudinage".

Interruption of bedding by burrowing animals would not explain the tension cracks, the pinching and swelling beds, and lack of alignment of gaps in the beds. Macrofossils are almost totally absent in the Ireton except for brachiopods and crinoids in the upper unit, but this does not entirely preclude the presence of soft-bodied burrowing animals.

SUMMARY

The following characteristics are typical of sedimentary "boudinage": (1) The "boudinage" structures should be coarser grained and hence less plastic at the time of deposition than the sediment surrounding them; (2) they may show evidence of tensional stresses such as fine peripheral cracks; (3) they may have thread-like connections; (4) nodules in the same plane should show similarities suggesting that they originated from the same bed and dissimilarities to the matrix; (5) there may be all gradations from undeformed laminations to isolated scattered nodules.

ACKNOWLEDGMENTS

The subject of this paper forms part of a thesis written under Professor J. Marvin Weller at the University of Chicago, to whom the writer is indebted for the careful reading of the manuscript and helpful suggestions. Discussions with Professors Hans Ramburg and John C. Jamieson were also very helpful. The writer wishes to thank Canadian Seaboard Oil Co. for releasing this material for publication.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 321–332
FIGS. 1–6, SEPTEMBER, 1958

DISTRIBUTION OF BRACHIOPOD AND PELECYPOD SHELLS BY CURRENTS[1]

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ABSTRACT

A statistical method of discriminating between life and death assemblages of brachiopods in fossil shell beds is demonstrated. The method utilizes the relative numbers of pedicle and brachial or right and left valves present in a sample of shell bed as well as the relative numbers of large and small shells and the relative numbers of articulated and disarticulated specimens.

Application of this method to a brachiopod-rich sample of a shell bed from strata of Early Devonian age in northern Maine shows that two of the contained species are indigenous whereas a third is not. The relationship between the number of articulated and disarticulated valves reveals that one of the two indigenous species has a unit disarticulation rate about eight times greater than the other. The extreme rarity of certain species in any one area is illustrated by the fact that in a sample of almost 5000 specimens, one species is represented by only one specimen.

Theoretical considerations of the transportation of articulated and disarticulated shells suggest that by contouring the amount of disarticulation of a species from a region it should be possible to locate the source area from which the shells came. It should also be possible to locate the source area by contouring the ratio of disarticulated opposite valves of a species from one region.

INTRODUCTION

In an earlier paper one of the authors has shown that statistical treatment of a sample of fossil shell bed can indicate whether it contains a life assemblage or a death assemblage (Boucot, 1953). Later, ways in which the experiment could have been usefully implemented became apparent, and a second statistical study of a sample of shell bed was made. This paper presents the results of this second study together with the theoretical considerations on the current distribution of bivalve shells that were an outgrowth of the statistical work.

The investigation was initiated by Boucot, and he is responsible for the interpretations of the resulting data. The actual measurement and preparation of the sample were done by DeMar in collaboration with Boucot. The statistical data were compiled by DeMar, and the relative rates of disarticulation for *Platyorthis* and *Leptocoelia* were calculated by Brace. The laboratory work was done under the auspices of the U. S. Geological Survey.

[1] Publication authorized by the Director, U. S. Geological Survey. Manuscript received February 1, 1958.

STATISTICAL STUDY OF A SAMPLE OF SHELL BED

Earlier statistical work with samples from brachiopod-rich shell beds (Boucot, 1953) had been limited to determination of size (width) and number of specimens belonging to each species. In the present investigation, records were kept of the number of whole shells, brachial valves, and pedicle valves of each species as well as of the width of each specimen.

Preparation of the Sample

The sample (U. S. Geological Survey Siluro-Devonian Fossil Loc. No. 2701) was a 50-pound block from a shell bed, about 6 in. thick, collected from strata of Oriskany (Early Devonian) age belonging to the Moose River Sandstone in a railroad cut about 1 mile southeast of Tarratine, Brassua Lake Quadrangle, Somerset County, Maine. The bivalves contained in the sample consisted almost entirely of brachiopods which were present as original calcareous shell material in a matrix of argillaceous sandstone (subgraywacke). Cleavage planes cut the bedding planes of the specimen at about a 30° angle.

The sample was divided into five blocks which were studied separately. Each block was weighed and then placed in a vat of di-

lute hydrochloric acid, where it remained until all exposed calcareous material had been dissolved. The block was then removed, rinsed, and soaked in a vat of water for several hours to remove all traces of acid and soluble chlorides from the interstices. After removal from this wash water the block was set aside to dry, usually for a period of about 12 hours.

The species and condition of shell (whole valve, pedicle valve, or brachial valve) of the brachiopod impressions exposed on each block were then identified, and the internal impression width of each specimen was

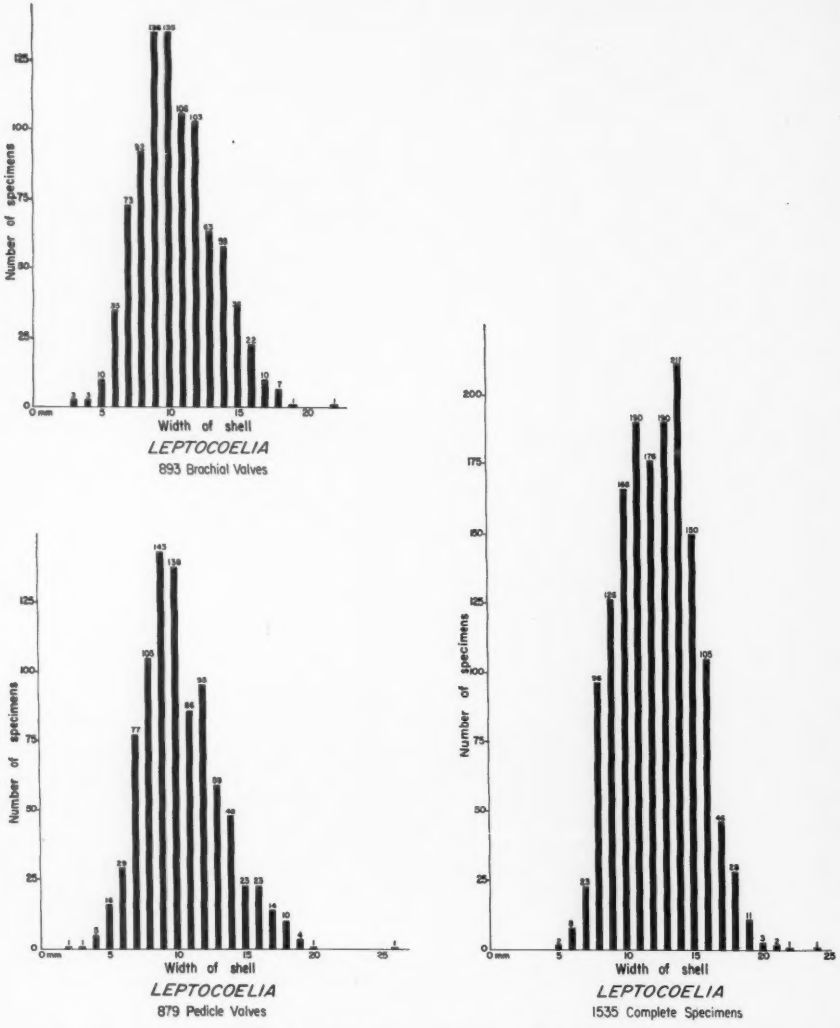


FIG. 1.—Size-frequency distribution of *Leptocoelia* from U.S.G.S. locality #SD-2701, Somerset Co., Maine.

measured to the nearest tenth of a millimeter with a vernier caliper. Measured specimens and their counterparts were marked with an indelible, acid-fast red pencil in order to prevent later remeasuring. Shells that were only partly visible, so that measurement of their full width was not possible, were fully exposed with the aid of a vibro-tool and dental tools. Specimens already measured were destroyed if they obscured partially exposed specimens.

After the limit of acid penetration had been reached, each block was split parallel to the bedding using a rock trimmer, and the entire process was repeated until it became impractical to split the material further.

Information Obtained from Sample

The frequency histograms (figs. 1–4) and the statistical data (tables 1–2) calculated for the collected information led to the following conclusions:

1.—The genera *Leptocoelia* and *Platyorthis* belong to a life assemblage that has not undergone significant transportation.

2.—The genus *Leptostrophia* is not an indigenous component of the *Leptocoelia-Platyorthis* life assemblage[2] at the locality from which this sample was collected and was probably brought to this site by outside current action.

3.—The remaining brachiopod genera in the sample, *Meristella*, *Beachia*, and *Chonetes*, are not represented by a large enough sample to provide an adequate basis for determining whether they belong to the *Leptocoelia-Platyorthis* life assemblage or to a death assemblage.

4.—*Platyorthis* in the sample disarticulated approximately eight times as readily as *Leptocoelia*.

5.—The competent velocity (the current velocity necessary to put the shells in motion) of articulated specimens of *Leptocoelia* is probably lower than that of disarticulated specimens of the same width.

6.—The competent velocity of disarticulated specimens of *Platyorthis* is probably lower than that of disarticulated shells of *Leptocoelia* of the same width.

7.—The competent velocity for the specimens of *Leptocoelia* and *Platyorthis* in the sample is greater than the current velocity that brought in the specimens of *Leptostrophia*.

8.—The information found for each of the five parts into which the sample was divided is not different from the total except in magnitude and the ratio of articulated to disarticulated specimens.

Discussion of Information Obtained from the Sample

The histograms and statistical information for the specimens of *Leptocoelia* (fig. 1.; table 1) in the sample show that the numbers of disarticulated pedicle and brachial valves are almost the same (879 pedicle valves; 893 brachial valves), and that the means and medians calculated for the disarticulated pedicle and brachial valves (table 1) are nearly equivalent. Because of these almost identical numbers of disarticulated valves of the same size and because of differences in shape (the brachial valve is relatively flat; the pedicle valve is highly convex and somewhat larger), it is most unlikely that the specimens in the sample can have been moved very much. The number of articulated *Leptocoelia* shells is 1535, considerably greater than the number of disarticulated specimens, and the mean is much greater. The greater mean suggests that small articulated specimens of this genus have a lower competent velocity than disarticulated shells of the same width.

The information for *Platyorthis* (fig. 2; table 1) in the sample shows that the number of disarticulated pedicle and brachial valves is almost identical (561 pedicle valves; 548 brachial valves) and that the means and medians calculated for the disarticulated valves (table 1) are nearly identical. Because of these similar numbers of disarticulated valves of the same size and because of differences in the shape (the brachial valve is relatively flat; the pedicle valve is highly convex and somewhat larger), it seems clear that the specimens in the sample have not been moved very much. The number of articulated shells of *Platyorthis* in the sample is very low (14), and the value of the mean, therefore, is not very re-

[2] The term life assemblage as used in this paper includes only that portion of the original life assemblage that was not removed by current action; that is, it is inferred that a large number of small specimens were removed by current action from the original life assemblage.

TABLE 1.—*Statistical data for the five blocks from U.S.G.S. locality No. SD-2701, Somerset Co., Maine*[1]

	Block No.					
	1	2	3	4	5	Total
Leptocoelia						
Pedicle						
Mean	9.88	11.06	9.68	10.79	10.33	10.36
Median	9.48	10.77	9.15	10.19	10.02	9.95
Standard d.	2.80	2.81	2.81	3.01	2.65	2.92
Percent	19.5	17.2	17.5	38.3	16.5	17.8
No. of specimens	182	258	136	31	272	879
Brachial						
Mean	10.23	11.02	9.60	10.50	10.65	10.48
Median	9.87	10.86	9.43	10.00	10.38	10.21
Standard d.	3.03	2.74	2.42	2.98	2.63	2.78
Percent	21.3	17.2	21.0	27.2	15.4	18.1
No. of specimens	199	258	161	22	253	893
Complete						
Mean	12.78	12.70	11.76	12.43	12.28	12.35
Median	12.68	12.73	11.48	12.33	12.28	12.39
Standard d.	3.13	2.77	2.54	1.68	2.58	2.75
Percent	17.3	39.5	17.7	8.6	38.7	31.2
No. of specimens	162	594	136	7	636	1535
Platyorthis						
Pedicle						
Mean	15.68	16.33	16.69	17.40	16.26	16.24
Median	15.87	16.33	17.00	17.72	15.77	16.44
Standard d.	6.68	5.48	5.42	3.83	5.55	5.76
Percent	10.9	11.5	11.6	6.2	11.7	11.4
No. of specimens	102	173	89	5	192	561
Brachial						
Mean	16.00	17.25	16.44	16.60	16.03	16.74
Median	16.14	18.62	16.78	17.72	16.54	17.01
Standard d.	5.32	4.97	5.90	5.71	5.72	5.52
Percent	12.6	10.1	12.0	6.2	11.0	11.1
No. of specimens	118	152	92	5	181	548
Complete						
Mean	26.00	16.33	21.25	—	17.00	20.57
Median	27.00	15.00	24.00	—	20.00	22.00
Standard d.	3.61	4.19	6.64	—	5.72	6.47
Percent	0.4	0.2	0.5	—	0.2	0.3
No. of specimens	4	3	4	—	3	14
Leptostrophia						
Pedicle						
Mean	21.85	24.70	20.90	19.50	22.33	21.82
Median	20.31	23.00	19.00	18.50	22.50	20.31
Standard d.	8.90	8.88	8.69	7.24	8.43	8.75
Percent	15.2	2.4	16.3	4.9	4.3	7.7
No. of specimens	142	37	125	4	70	378
Brachial						
Mean	18.90	18.20	19.00	—	22.86	19.63
Median	20.33	19.00	19.00	—	25.67	19.81
Standard d.	5.19	6.34	3.28	—	4.36	4.89
Percent	1.1	0.3	1.7	—	0.4	0.7
No. of specimens	10	5	13	—	7	35

	Block No.					
	1	2	3	4	5	Total
Chonetes						
Pedicle						
Mean	25.70	29.58	31.17	34.67	26.87	28.20
Median	30.00	30.50	30.50	40.00	26.50	30.00
Standard d.	12.70	8.32	9.86	9.69	10.33	10.81
Percent	1.4	0.8	0.8	3.7	1.0	1.0
No. of specimens	13	12	6	3	16	50
Brachial						
Mean	39.00	31.44	31.60	33.67	26.36	30.50
Median	39.00	33.00	34.00	34.00	24.00	33.50
Standard d.	5.00	6.94	7.05	3.45	11.09	9.30
Percent	0.2	0.7	0.7	3.7	0.7	0.6
No. of specimens	2	11	5	3	11	32

[1] All measurements given in mm

liable. The relatively low number of articulated specimens of *Platyorthis* relative to *Leptocoelia* suggests that *Platyorthis* disarticulates at a far greater rate than *Leptocoelia*. This conclusion is further substantiated by the observations of Boucot that articulated specimens of *Platyorthis* in other regions are rare, whereas those of *Leptocoelia* are common.

The medians for disarticulated *Leptocoelia* are much lower than those for disarticulated *Platyorthis* (table 1). The competent velocity for disarticulated specimens of *Platyorthis*, therefore, should be much lower than that for disarticulated specimens of *Leptocoelia*.

Leptostrophia, the third brachiopod genus that is present in significant numbers in the shell bed (fig. 2), is concluded to have undergone extensive transportation because (1) there are 378 pedicle valves in the samples, but only 35 brachial valves, and (2) the opposing valves of *Leptostrophia* are almost mirror images as well as being bilaterally symmetrical and so could be expected to have very similar hydrodynamic behavior. The widely differing means and frequency for the pedicle and brachial valves of *Leptostrophia* in this sample (table 1) support the conclusion that extensive separation of pedicle from brachial valves had taken place during transportation. In view of the strong evidence that the accompanying specimens of *Leptocoelia* and *Platyorthis* have not moved very much and belong, therefore, to an indigenous life assemblage, it is unlikely that *Leptostrophia* could also have been an indigenous form whose brachial valves had been selectively winnowed out by current action. It follows that the current velocity necessary to keep disarticulated valves of *Leptostrophia* in motion is lower than the competent velocity of the associated specimens of *Platyorthis* and *Leptocoelia* and of the associated sediment. Boucot has observed that brachial valves of *Leptostrophia* are very rare, both in the field and in museum collections, a finding which supports the findings of this experiment. The comparative rarity of *Leptostrophia* brachial valves raises the difficult question of their ultimate disposition in view of their morphological similarity to the relatively common pedicle valves.

Because of the small numbers of specimens found in the shell bed, significant conclusions can not be made whether *Chonetes*, *Beachia*, and *Meristella* belong to the life assemblage or the death assemblage. However, it is notable that in more than 5000 specimens counted, there were only one *Meristella* and three *Beachia;* the latter is the critical dating element.

Rates of Disarticulation

Relative speeds of disarticulation of *Platyorthis* and *Leptocoelia* can be calculated, based on analysis of the abundance of whole and disarticulated specimens in the four blocks which were originally side by side in the shell bed. To do this it is first necessary to examine the disarticulation process more closely and to find a simple mathematical relationship to describe this process.

Disarticulation can be assumed to in-

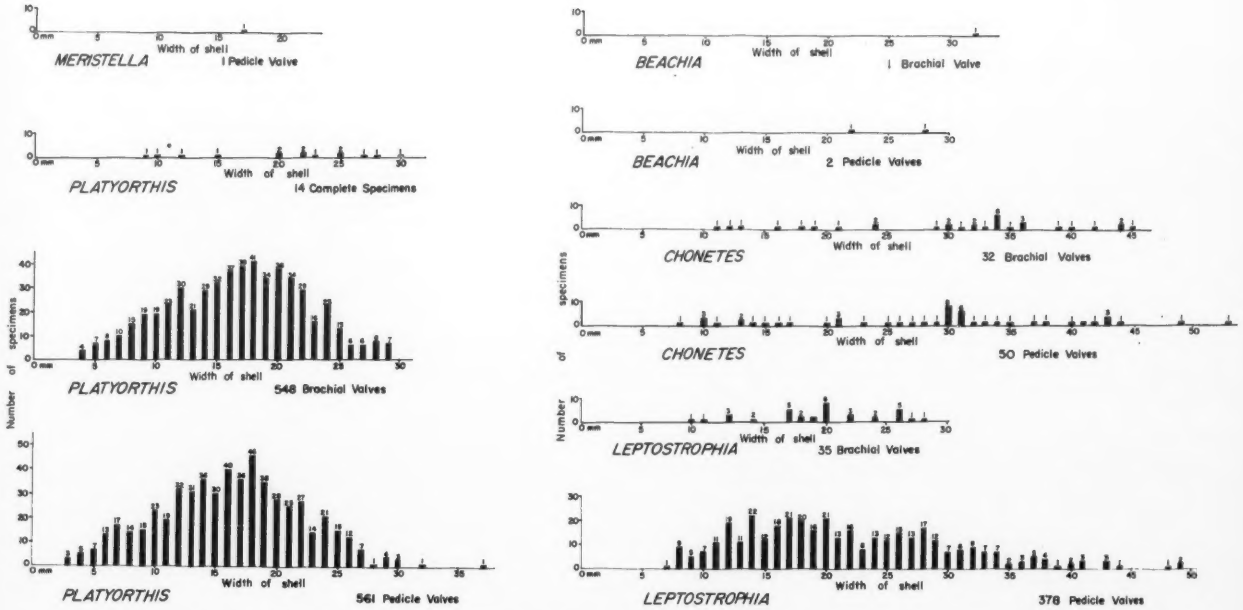


FIG. 2.—Size-frequency distribution of brachiopods other than *Leptocoelia* from U.S.G.S. locality #SD-2701, Somerset Co., Maine.

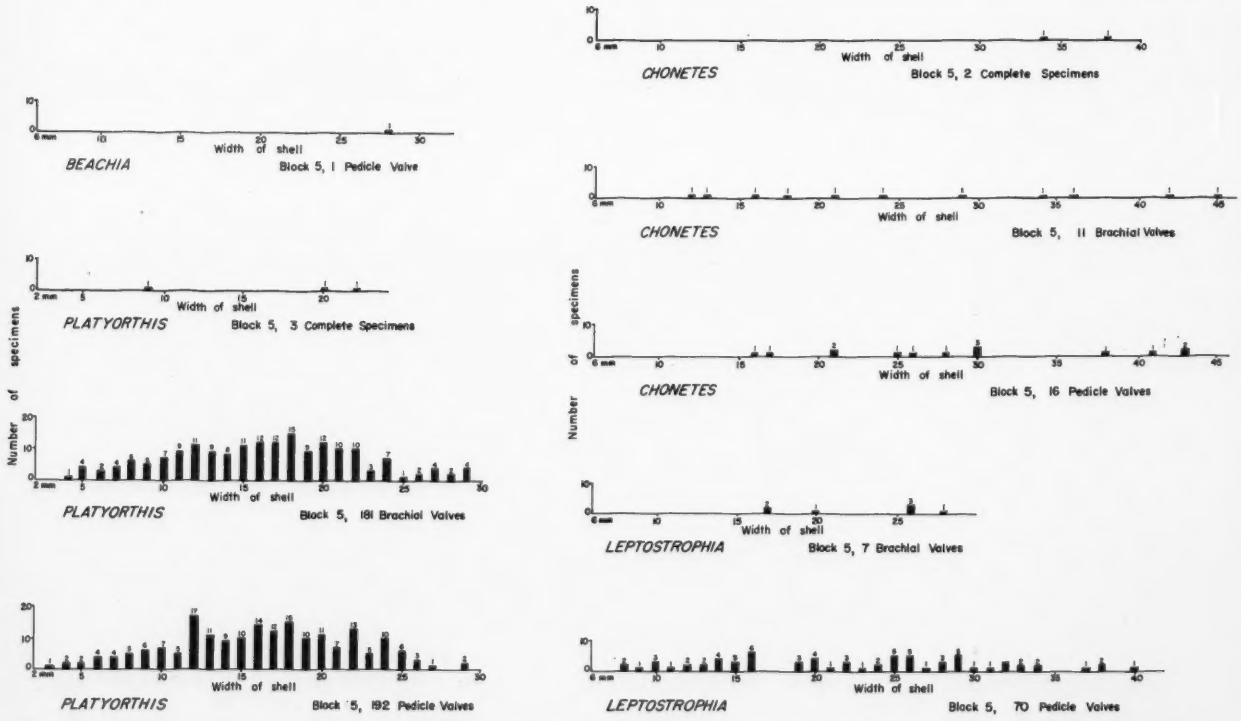


FIG. 3.—Size-frequency distribution of brachiopods other than *Leptocoelia* in block 5 from U.S.G.S. locality #SD-2701, Somerset Co., Maine.

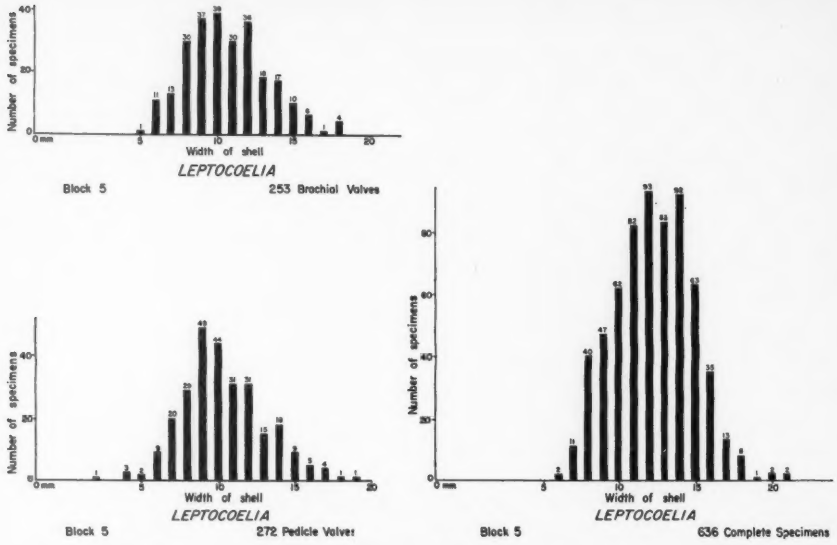


FIG. 4.—Size-frequency distribution of *Leptocoelia* in block 5 from U.S.G.S. locality #SD-2701, Somerset Co., Maine.

crease solely with the total distance, D, that a sample of shells moves from the place where the animals died. D may or may not be equal to d, the net distance from the death site to final resting place in the shell bed. A first simple assumption is that disarticulation follows an exponential law. That is, that the decrease in the number of articulated shells, N, with distance, D, is directly proportional to the number of whole shells present at any one time in the assemblage. Or, in equation form, that $dN/dD = -kN$, where k can be called the *disarticulation constant.* The usual integrated form of this expression is $N/N_0 = e^{-kD}$, where e is 2.718 and N_0 is the original number of whole shells present before disarticulation. N/N_0 will be called the *disarticulation ratio.*

In the above equation there will probably be a different k for each species. Without knowing actual distances k cannot be calculated; however, with the use of the data in table 2 it is possible to compute the ratio of k of *Platyorthis* to k of *Leptocoelia.* As mentioned previously, these two genera are concluded to be part of a life assemblage that has not been transported away from

TABLE 2.—*Number of whole and disarticulated Platyorthis and Leptocoelia shells in four blocks from sample SD-2701*

Block No.	*Platyorthis*				*Leptocoelia*			
	Articulated		Disarticulated	Total	Articulated		Disarticulated	Total
	No.	%	No.	No.	No.	%	No.	No.
1	4	3.5	110	114	162	46	190	352
2	3	1.8	163	166	594	70	258	852
3	4	4.2	91	95	136	48	149	285
4	3	1.6	186	189	636	71	263	899
Total	14			564	1528			2388

the source area and, therefore, D is the same for both.

The number of whole and disarticulated specimens of *Platyorthis* and *Leptocoelia* present in four adjacent blocks is shown in columns 2, 4, 6, and 8 of table 2. The original number of articulated shells, N_0 (columns 5 and 9) is the sum of articulated and disarticulated specimens now present in the sample. The number of disarticulated shells (columns 4 and 8) is the average of the numbers of pedicle and brachial valves found; in all four samples the difference in these two numbers was less than 8 percent.

The disarticulation ratio for *Platyorthis* is

$$\frac{14}{564} = 0.025$$

and for *Leptocoelia* is

$$\frac{1528}{2388} = 0.64.$$

The ratio of k's will then be

$$\frac{k_{Platy.}}{k_{Lept.}} = \frac{\ln 0.025}{\ln 0.63} = 8.3.$$

In an easily visualized form, this ratio simply tells us that the distance required to disarticulate the original number by a given percent for *Leptocoelia* is 8 times that for *Platyorthis*.

The disarticulation constant, k, is assumed here to remain constant. A test of this assumption would be to study similar abundance data elsewhere for *Platyorthis* and *Leptocoelia*. Flume studies with shells of various shapes would of course enable more exact relationships to be determined and could ultimately be of great value if it is found that disarticulation behavior of large, geometrically similar groups of shells is the same.

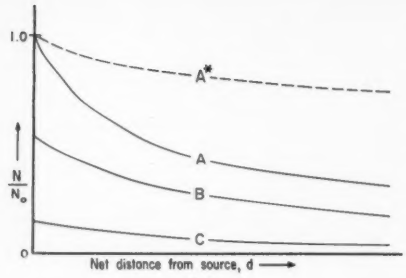
Locating the Source Area

An obvious extension of the above treatment of disarticulation is to the location of the source area of an assemblage of shells. It should simply be a matter of extrapolating from localities with various values of N/N_0 to the point at which N/N_0 is equal to 1. However, care is necessary here in distinguishing between the distances D and d as defined earlier. To and fro motion, which may disarticulate a shell, may not appear in the net distance, d, which the shell has apparently moved. Some shell samples will be almost completely disarticulated by oscillatory currents but remain more or less at the source area (*Platyorthis* in Sample No. SD-2701). In figure 5a a number of plots are shown of the theoretical variation of N/N_0 with d, the net distance from the source. Curves A, B and C are for species with a constant, k, different from that of another species represented by curve A^*. Curve A is the ideal situation in which articulated shells have moved in a uniform direction away from the source area and were disarticulated at a rate which was large in terms of the distances involved. Curve C represents a situation in which disarticulation occurred largely by to and fro motion in place; the sample was almost completely disarticulated before being transported from the source. Curve B represents a situation (*Leptocoelia* in Sample No. SD-2701) intermediate between A and C. Curve A^* differs from A in that the rate of disarticulation is smaller in terms of the distance (d) involved.

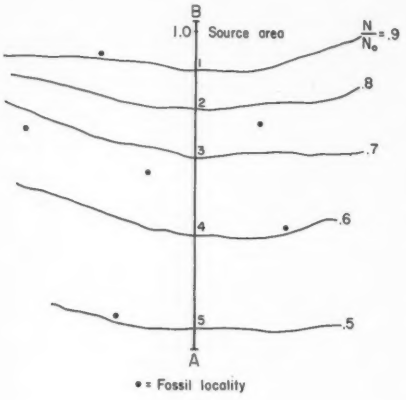
In order to locate the source area it is necessary to extrapolate back in distance to the point where $N/N_0 = 1$. This can only be done if D is equal to d. Other evidence from the shell bed must be available to show that simple to and fro motion leading to disarticulation was essentially absent.

Supposing that the ratio N/N_0 was known for a species at a number of places, or ideally, that one could prepare contours of N/N_0 over an area, as shown in figure 5b. The point at which N/N_0 becomes equal to one is found as follows: First, a line AB is laid off as nearly perpendicular to the contours as possible. Distances along AB and the values of N/N_0 are next plotted as in figure 5c and a curve drawn through the points. At the value on the curve where N/N_0 equals one, $Y - Y'$ is drawn; the distance $Y' - 1$ locates the source. In the example shown in figure 5c, CC' resembles curve A^* of figure 5a. CC' is an exponential curve and the extrapolation should be carried out on this basis. If the measured points do not fall on an exponential curve then the original assumption as to variation of disarticulation with distance will have to be modified accordingly.

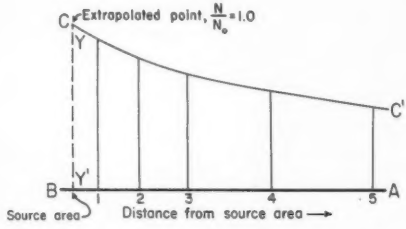
In those instances for which the change in N/N_0 with d fall in the class of curves



(a)



(b)



(c)

FIG. 5. a—Theoretical variation of disarticulation ration N/N_0 with net distance from source. See text for explanation.

b—A hypothetical map area showing a number of fossil localities at which the ratio N/N_0 is measured for some species. N is the remaining number of articulated shells, N_0 the number of original shells. Line AB is roughly perpendicular to contours which are drawn through equal values of N/N_0.

c—A plot showing distances between contoured values of N/N_0 (such as 1–2 in (b)) against values of N/N_0 on the vertical axis. An exponential curve, C-C', is fitted to the points and extended to the point, Y, where N/N_0 equals 1. The derived distance $Y'-1$ locates the source area as indicated in (b).

typified by curve B of figure 5a it will not be possible to determine from ratios of N/N_0 alone whether or not the source area has been located. However, by contouring ratios of pedicle to brachial valves for brachiopods and left to right valves for pelecypods it will be possible to qualitatively check the location of the source area. In those instances where essentially to and fro motion in the source area has resulted in large scale disarticulation the ratio of opposite valves of the same width will remain equal to unity. Transportation away from the source area will result in sorting due to differences in competent velocity between the opposite valves when subjected to a succession of currents with differing velocities, it being very unlikely that over a period of time an area will be swept by currents with velocities always greater than the competent velocity of both valves. In addition, it is clear that opposite valves having different hydrodynamic characteristics will be deposited separately by a waning current. For example, from consideration of sample SD-2701 it is clear that the ratios of pedicle to brachial valves for *Leptocoelia* and *Platyorthis* are practically unity, whereas N/N_0 departs greatly from unity, leading to the conclusion that the shells belonging to these two genera have not been transported a great distance (d) but have been subjected to disarticulating movements (D). In the case of *Leptostrophia* from sample SD-2701 the 10:1 ratio of pedicle to brachial valves suggests that large scale transportation (d) has taken place before eventual deposition at the final burial site.

CONCLUSIONS

The preceding study has shown that the statistical approach can indicate which components of a shell-bed fauna are indigenous and which are not. After the paleoecologist has determined which components of his fauna belong to a death assemblage (see also Boucot, 1953, p. 25–40), he can then consider (1) whether these components have a common source or more than one source, and (2) how far the components have trav-

elled, and in what direction or directions they have travelled.

Such consideration is useful because it can provide the geologist with information regarding the vectorial current direction[3] in the area and stratum studied and the paleoecologist with information regarding the source area and life assemblages represented by the death assemblages studied.

Initial Assumptions

In considering the problem of current distribution of brachiopods and pelecypods and in dealing with the statistical shell bed study, certain initial assumptions have been made:

1.—The vectorial current direction determines the final resting place of the shells.

2.—After the decomposition of their soft parts, dead brachiopod and pelecypod shells tend to become disarticulated into their component valves by current action.

3.—Each brachiopod and pelecypod species has its own rate of disarticulation; this rate probably varies within each species depending on the size of the specimens. The degree of disarticulation that a population shows is a function of the rates of disarticulation of its components plus the duration and magnitude of the disarticulating forces. The chief reason for assuming that different species have different rates of disarticulation is that some species are frequently found in an articulated condition whereas others are not. Boucot (1953, p. 32, 39) discusses several examples taken from the Lower Devonian of northern Maine, and every paleontologist can recall examples of this phenomenon observed in the course of his own work. The statistical work described above has shown how the relative difference in rate of disarticulation for two associated species may be determined.

4.—In general, whole shells and disarticulated valves of the same species will have differing competent velocities. This assumption is partly supported by the data of Menard and Boucot (1951, p. 140–143, fig. 3) who found different competent velocities for articulated and disarticulated shells of the same species in the course of their flume experiments on the movement of brachiopods by water (fig. 6).

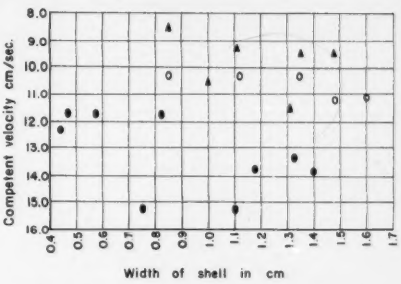


FIG. 6.—Diagram showing competent velocity of whole shells (black dots), brachial valves (triangles), and pedicle valves (circles) of *Terebratulina septentrionalis*. Information from Menard and Boucot, 1951, table 1.

5.—The disarticulated pedicle and brachial valves of a brachiopod, being of different size and shape, will have different competent velocities which will tend to disperse them during transportation. Because they are bilaterally symmetrical, these valves will move in a direction parallel to that of the transporting current. The opposing valves of a pelecypod, on the other hand, will tend to have identical competent velocities since the shells are mirror images. But because they are also asymmetric, they will tend to travel in directions that diverge on either side of the current direction. Martin-Kaye (1951) has found a tendency for right valves to be deposited on some parts of a beach and for left valves to be deposited on other parts of the same beach. The divergent movement of mirror image pelecypod valves might be compared to the port or starboard tack of a sailboat.

Changes in Population Character During Transportation

Whether the exodus of a mass of dead shells from its source area is almost instantaneous owing to a catastrophe or whether it is prolonged over a period of years, the character of the finally interred population or populations will be unchanged except for population density. The population density will decrease if the rate of sedimentation is high, as previously found by Eagar (1952, p. 359–361). In the simplest case, the source area will be swept by currents whose direc-

[3] "Vectorial current direction" is used here as the net direction and distance from the source area to the area of deposition that a population of shells has traversed after death.

tion is unvarying. Whole shells of the same size will travel from the source area with the current at a velocity proportional to that of the transporting current. Both their radial and transverse distribution[4] from the source area should be normal, reflecting variations of the shells from perfect bilateral symmetry as well as differences in size. The pedicle and brachial valves produced by the disarticulation of the whole shells as they are being transported will probably have a highly skew radial distribution and a normal transverse distribution, reflecting the rate of disarticulation and the degree of departure from perfect bilateral symmetry. In those cases where the rate of disarticulation is very high, that is, where the whole shells are disarticulated almost instantaneously upon being subjected to disarticulating forces, the radial distributions should be normal rather than skew. Disarticulated mirror-image shells will have a normal radial and transverse distribution, but the radii will not correspond to the current direction due to the dextral and sinistral divergence of the shells.

It should be possible to determine the general location of the source area for a fauna of brachiopods now found in one stratum by contouring the amount of disarticulation, that is, by measuring and contouring the ratio of disarticulated to whole shells over as large an area as possible. The direction of the source should be normal to the contours of the disarticulation ratio and those of the ratios of opposite valves. The distance back to the source area can be determined graphically by plotting known values of N/N_0, the distances between them, followed by an extrapolation to the distance where $N/N_0 = 1$. The only assumptions here are that disarticulation of the shells changes exponentially with distance (D) and that current flow was the only force acting to disarticulate the shells.

Asymmetric mirror-image shells, as mentioned previously, will tend to diverge dextrally and sinistrally from the direction of the currrent. If the rate of disarticulation is high, so that the shells disarticulate almost instantaneously, then the angular divergence of the bulk of the original sample would be initiated at once. By contouring the frequency of right and left valves and drawing lines through the elongate directions of greatest frequency, one should obtain an intersection corresponding to the source bed. The bixectrix of the angle between the lines would be parallel to the direction of the current that moved the shells.

[4] Radial distribution is the size-frequency distribution to be found along tracks emanating from the source area. Transverse distribution is the size-frequency distribution along lines normal to these tracks.

ACKNOWLEDGEMENTS

It is a pleasure to acknowledge the interested comments and criticisms of Drs. S. H. Straw and F. M. Broadhurst, Dept. of Geology, The University, Manchester, Drs. S. O. Schlanger and E. L. Yochelson, U. S. Geological Survey, Washington, D.C., and Mr. Richard Maehl, Department of Geology and Geophysics, Massachusetts Institute of Technology, Cambridge, Massachusetts.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 333–341
FIGS. 1–10, SEPTEMBER, 1958

RECENT GROOVING IN LAKE BOTTOM SEDIMENTS AT GREAT SLAVE LAKE, NORTHWEST TERRITORIES[1]

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ABSTRACT

Large grooves up to 33 m wide and 4760 m long have been inscribed in soft unconsolidated lake bottom sediments at Great Slave Lake, NWT. The grooves are probably formed by wind-driven ice floe gouging during each spring breakup. Smaller grooves having a boulder at one end were formed by boulders pushed by floating lake ice. Study of the hydraulics of fluid flow upon spheroidal solids indicates that wind is quite incapable of transporting large rocks like those suggested to have formed the grooves on Racetrack Playa, California.

INTRODUCTION

Attention has been drawn to the spectacular grooves inscribed in the playa lake bottoms in California by McAllister and Agnew (1948), Clements (1952), Kirk (1952 a, b), Shelton (1953), and Stanley (1955). Boulders up to 274 kilograms in weight are often found at the ends of the grooves and the early writers invoked wind as the motivating force. Stanley (1955) recorded the "signatures" of numerous boulders and suggested that the grooves were formed by boulders entrapped in ice floes formed from the break-up of ice sheets, which were moved by wind on the shallow water lakes.

The Great Slave Lake region, NWT, provides excellent examples, not only of large grooves scribed by glacial erratics but also of enormous furrows produced by the ice floes themselves.

THE GREAT SLAVE LAKE AREA

The grooves are best developed in the very shallow waters at the extreme western end of the southern part of Great Slave Lake where the Mackenzie River leaves the lake by way of a large outlet choked with islands and sand bars (fig. 1).

The localities studied included North Channel (61°08′–61°13′ N, 116°40′–117°00′ W), Deep Bay (61°13′–61°19′ N, 116°35′–116°48′ W) and Falaise Lake (61°28′ N, 116°10′ W). Large grooves are also seen in the area northwest of North Channel (centered at 61°20′ N, 117°00′ W). The region is one of very low relief, only slightly above the level of Great Slave Lake. The presence of raised beaches indicates that it was part of the lake bottom during Pleistocene or later times. Swampy muskeg comprises the plant cover and peat accumulations are numerous.

DESCRIPTION OF THE GROOVES

Measurements on aerial photographs show that grooves may range from small furrows a few meters long to channelings up to 33 m wide and 4760 m long. Figure 2 illustrates the extraordinary length of some of the grooves. Where seen on the ground, at Falaise Lake, the grooves measure up to 1 meter deep.

A striking feature of these grooves is their linearity. Few are irregular and none are zig-zagged. Slight curvature, amounting to a few degrees, however, is commonly exhibited by the larger grooves.

In the North Channel, a general trend of 330° is evident although a range of lineation directions from 305° to 355° occurs. It appears that the largest grooves are directed about 345 to 350°. In some cases (fig. 3), a set of criss-crossing grooves intersecting at angles of 45 to 55 degrees forms a braided pattern.

In all cases the grooves are inscribed in unconsolidated lake bottom sediments. At North Channel, soft muds form the river floor and shores, whereas at Falaise Lake, the grooves contain gravel. The Simpson Shales (Upper Devonian) are suspected to underlie the area but no exposures appear in the vicinity. The grooves appear to bear no

[1] Manuscript received April 18, 1958.

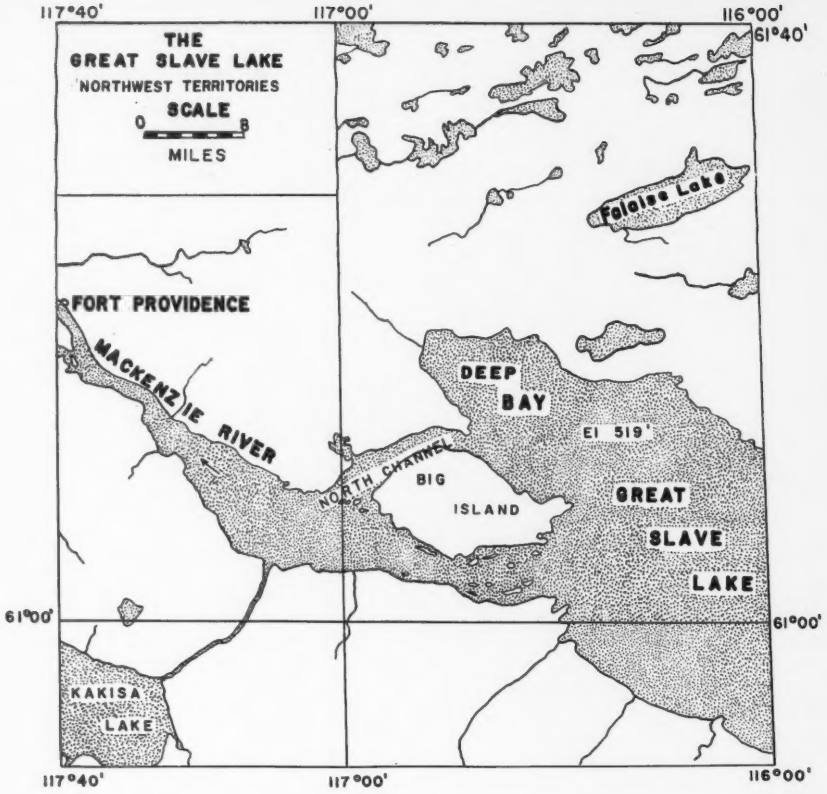

FIG. 1.—The Great Slave Lake area, Northwest Territories.

relationship to bedrock which is probably covered by many meters or tens of meters of river and lake alluvium.

The lowland area northwest of North Channel, now covered by muskeg swamps and peat accumulations can be shown by evidence of raised beaches to have once been under the waters of Great Slave Lake. This area is also extensively grooved (fig. 4) but gradual obliteration by erosion makes the grooves less conspicuous.

Because at Falaise Lake the grooves truncate pronounced glacial lineation, they are recognized as Post-Pleistocene in age. The truncation is illustrated in figure 5.

Many of the smaller grooves, less than 30 meters long and a meter or so wide contain a boulder or glacial erratic at the end nearest to shore. Boulder grooves are especially numerous at Barlett Lake (63°08′ N, 118°20′ W), but are only of minor importance at North Channel and Falaise Lake. Figure 6 shows a groove of the order of 30 metres long with a boulder at the shoreward termination.

Figures 7 to 9 are photographs taken from a helicopter illustrating the grooves at Falaise Lake.

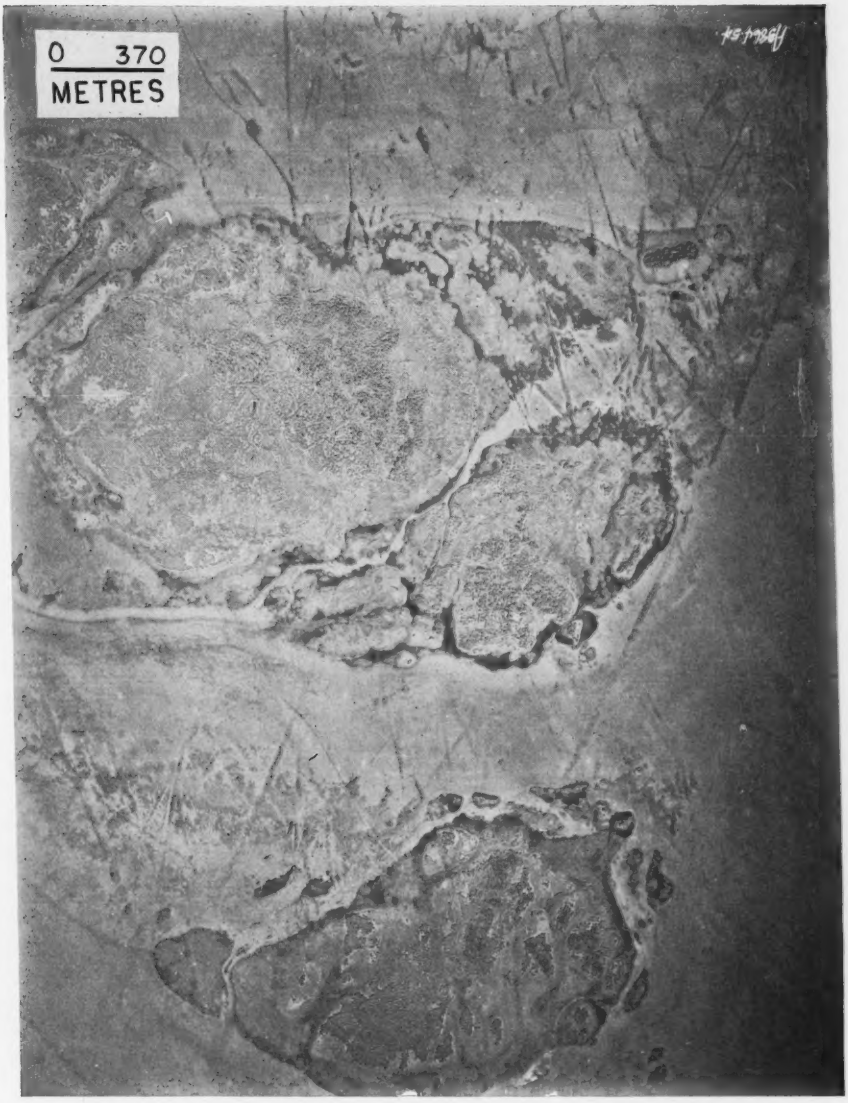

FIG. 2.—Grooves near North Channel showing extraordinary size. Courtesy R.C.A.F.

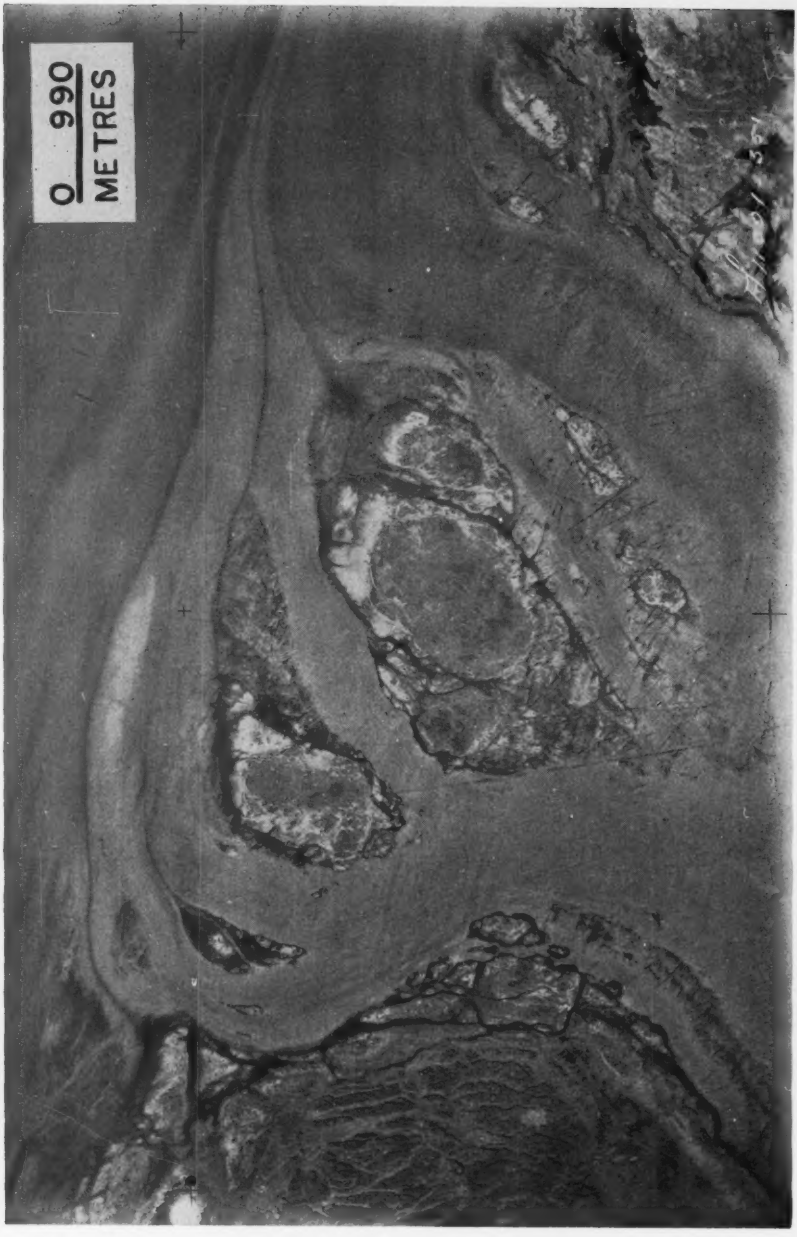

FIG. 3.—Criss-crossing pattern shown by grooves at North Channel. The area shown is about 53 sq. km. Courtesy R.C.A.F.

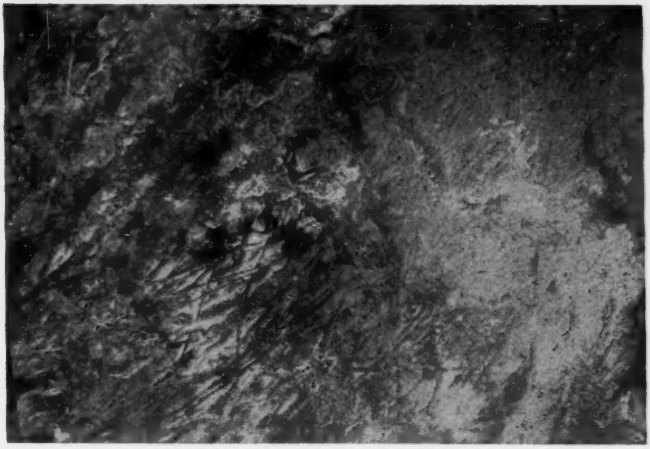

FIG. 4.—High altitude vertical view of the partly obliterated grooves in the area northwest of North Channel. The area was once part of Great Slave Lake.

FIG. 5.—Falaise Lake showing two large promontories created by glacial grooving of the Pleistocene Ice sheet. The smaller lake ice grooves are seen truncating the promontories at nearly right angles. Oblique aerial view looking west.

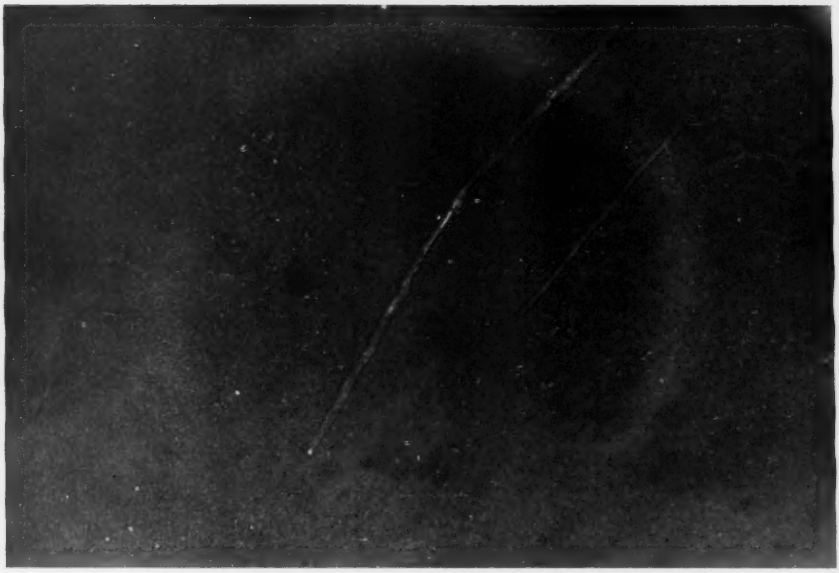

FIG. 6.—Groove with boulder at shoreward end. Of the order of 30 meters long. Vertical aerial view taken by helicopter of grooves in the very shallow water of Great Slave Lake.

FIG. 7.—High altitude view of Falaise Lake. The lake margin coincides with a horizontal line just above the middle of the photograph. Above this line is shown a large expanse of muskeg with darker areas of forest. In the foreground is the dried-up portion of Falaise Lake showing large grooves and smaller patches or pools of water.

FIG. 8.—Large grooves along the shallow-water shoreline of Falaise Lake.

FIG. 9.—Large grooves on a dried-up portion of Falaise Lake. Oblique aerial view. Scale is indicated by trees along shore.

DISCUSSION OF ORIGIN

Distinction from Glacial Furrowing

Furrowing and grooving caused by the last movement of the Pleistocene ice sheet in the Great Slave area are easily recognized by their size and constancy of direction.

Pronounced glacial furrows, many hundreds and even thousands of meters wide and several kilometers long are common, and many are occupied by small shallow lakes. At Falaise Lake, the crests of the undulations form long narrow peninsulas, whereas the troughs create elongate bays.

The direction of the glacial grooving is constant at approximately 070° in this region.

The cross-cutting relationships of glacial furrows and the grooves described in this paper are seen at Falaise Lake where the smaller grooves are inscribed in the low-lying peninsulas and bays cut by glacial ice (figure 5), the lineation of the smaller grooves being nearly perpendicular to that of the glacial grooves.

Origin by Boulder Movement

Evidence that some grooves were caused by boulders is the presence of a boulder at the termination of the groove. Such were seen in all localities, but the grooves are never more than a meter or so wide and never more than 30 m long. The absence of boulders at the ends of the larger grooves supports the view that boulder movement in the soft unconsolidated lake sediment apparently did not form them.

Wind and the Movement of Boulders

After the descriptions of trails on playa lake bottoms of California by Kirk (1952a), some controversy arose as to whether or not wind is capable of moving boulders over the slippery mud floor of a playa. Boulders up to 274 kilograms were reported by Kirk and winds in the area were seldom known to exceed 40 mph.

The hydraulics of the problem have been examined. Preston (personal communication 1958) has shown that the force exerted by a fluid upon a spheroidal solid assuming little or no turbulence in the "wake" area is of the order of $0.3\rho V^2l^2$ where ρ is the density of air, V is the velocity of the wind and l is the diameter of the solid. The force produced by the wind is equated with the force required to overcome static friction. This force is $\mu_s m$ where μ_s is the static coefficient of friction and m is the mass of the solid. The static coefficient of friction is taken from standard tables as 0.3 which is that of rock on wet earth. Assuming the specific gravity of the spheroid to be 2.8 or that of the average rock, and using the formula for the volume of a sphere, the variation of mass with the diameter can be derived. This can be substituted in the equation of the two forces to give the variation of the mass of the spheroid (which the wind of velocity V can move) with the velocity of the wind. The relationship is: $m = 8 \times 10^{-19} V^6$ where m is in grams and V is in cm/sec.

Figure 10 shows that the wind is capable of exerting high forces only when moving at very high velocities, but is quite incapable of moving boulders of the sizes reported from the Californian playas. Winds of 40 mph can only move boulders of the order of 25 grams and not 274 kilograms as suggested by Kirk (1952a).

Origin by Ice Floe Movement under the Influence of Wind

Stanley (1955), working with playa stone trails at Inyo County, California, has mapped the "signatures" of numerous boulders and by correlation of identical "signatures" was able to show that ice floes floating in the very shallow water had used enclosed boulders and rocks to inscribe the soft lake bottom.

A different argument is put forth to account for the grooves at Great Slave Lake in that boulders, because of their general absence, played only a minor role.

Ice floes broken off from the thick ice cover formed during the winter, and driven shoreward by the prevailing winds are the only objects capable of creating grooves up to 33 m wide.

In some of the larger lakes, earth ridges up to 1½ m high and parallel to the shoreline are composed of unconsolidated lake bottom sediments and are caused by shoreward ice movement. Occasionally trees near shore are uprooted. At Barlett and Windflower Lakes, several successive ridges could be distinguished.

The weather conditions prevailing in the Great Slave Lake area (Department of

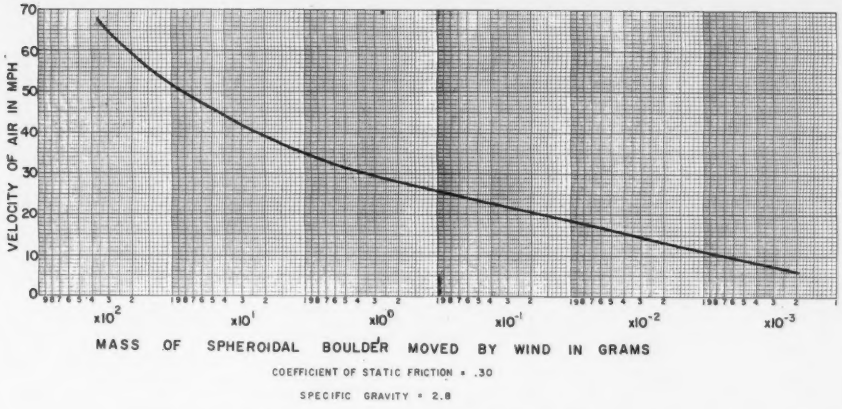

FIG. 10.—Relationship between wind velocity and the mass of a boulder moved by the wind over playa bottoms.

Transport 1954) substantiate this view. Breakup generally occurs in the month of May. During this month, the prevailing wind direction (measured at Yellowknife, NWT) is northward with a maximum wind speed of 26 mph. In April conditions are the same, but in June the wind direction varies considerably. Thus it is seen that the general trend of the grooves is correlative with that of the prevailing wind during the ice breakup.

CONCLUSION

During May and early June, the winter ice sheet formed over Great Slave and other lakes nearby, begins to disintegrate into large ice floes. These floes, with any boulders frozen into the ice in the shallow waters near shore, are driven shoreward by the prevailing winds, cutting large furrows in the soft lake bottom sediment.

ACKNOWLEDGEMENTS

The writer is grateful to Shell Oil Company of Canada, Ltd., for release of the photographs and observations made during a 1957 field survey and to Dr. M. A. Preston of McMaster University for derivation of the formula for the effect of a fluid on a spheroidal object.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 342–344
FIG. 1, SEPTEMBER, 1958

POSSIBLE ABANDONED BARRIER ISLANDS IN PANHANDLE FLORIDA[1]

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ABSTRACT

Two arcuate sand bodies, located in a swamp southwest of Tallahassee, were studied and compared with several shoreline features. It is concluded that they are probably abandoned, immature, barrier islands.

MacNeil (1949) presented a map showing the Pleistocene shore lines of Florida. He drew on this map, in Franklin County (southwest of Tallahassee) an arcuate string of low hills. In his accompanying text he suggested that this arc might have had an origin similar to that of the arcuate chain of islands off of the mouth of the Suwannee River. The present study was undertaken in an effort to check this suggestion.

The arcuate trend shown on MacNeil's map appears both in the field and on the air photos as a long, broad, low gently curving ridge. There is nothing to indicate that it is or was a string of discrete hills. This ridge is located in Tates Hell Swamp, about 14 miles northeast of Apalachicola, Florida. About two miles north of the first arcuate ridge is a second, similar, ridge. A third arc occurs about eight miles west of Apalachicola (fig. 1). Other less well defined arcuate ridges may be found in the same general area.

All three sharply delineated ridges trend east-west and are concave toward the north. The shortest of the three is about six miles long, the longest about eight. The width varies in general between 500 and 1500 feet; the relief, between two and 12 feet. U.S.G.S. air photos, flown in 1942 by Aero Service Corp., show all three arcs clearly. The contract symbol for this photography is GS-AB. Photos showing the western arc are 1–116 to 1–120, inclusive; the two eastern arcs appear on photos 1–50 to 1–54, inclusive. The two eastern arcs lie within the area covered by the Fort Gadsden and Tates Hell Swamp quadrangles of the U.S.G.S. These two arcs stand at elevations of approximately 20 and 30 feet, respectively, above sea level.

Field work was carried out by Brenneman and various assistants in the winter and spring of 1956 (Brenneman, 1957).

It was originally hoped that a comparison of the arcuate ridges with the arcuate chain of islands off of the mouth of the Suwannee River might be fruitful. Preliminary field work in the two areas, however, revealed that the ridges are made of quartz sand, whereas the islands are growing oyster "reefs." It was then decided to compare the ridges with other known sand features in an effort to determine their origin. For this purpose, detailed studies were made on the two arcs in Tates Hell Swamp, on a barrier island (St. Georges Island) about 12 miles south of the ridges, and on St. Joseph Spit, about 15 miles west of Apalachicola. The authors also considered the following types

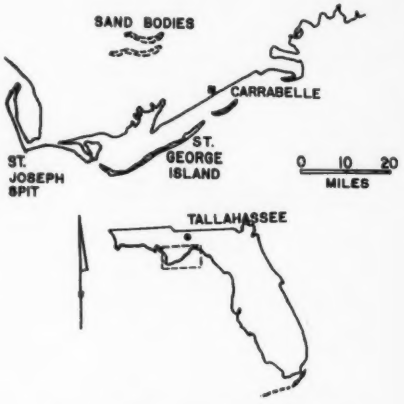

FIG. 1.—Map showing location of the two arcuate sand bodies studied.

[1] Based on a thesis submitted by Brenneman in partial fulfillment of the degree, Master of Science, at Florida State University. Manuscript received March 17, 1958.

TABLE 1.—*Phi median and phi decile deviation for four of the sand features studied*

	Md_ϕ	Decile deviation, $[D_9-D_1]/2.6$		(So (Mean)
		(Mean)	(Range)	
North sand body	2.45	0.93	0.60 to 1.29	1.49
South sand body	2.05	1.06	0.73 1.55	1.60
St. George Island	2.02	0.45	0.29 0.71	1.20
St. Joseph Spit	2.08	0.36	0.28 0.59	1.19

of features: sand dune, chenier or beach ridge, natural levee, river channel, dwip, lunate bar, and low-and-ball.

An erosional origin was rejected on the basis of both field work and air photo studies. The ridges in Tates Hell Swamp rise above an otherwise essentially featureless sand plain. If a scarp is associated with either ridge it is a small one located on the seaward side (south side).

The lunate bar and dwip (Bates, 1953) may be eliminated because of their much sharper curvature. The dwip is also convex landward whereas the sand ridges are convex seaward.

The low-and-ball of Evans (1940) is a much narrower feature, the ball being only 10 to 20 feet across.

The ridges are neither natural levees nor old sand-filled channels which now stand high because of the compaction of adjacent clays. The adjacent swamps are underlain by the same quartz sand which makes up the arcs. Further, the obvious east-west trend does not fit the regional north-to-south stream gradient which has persisted throughout Pleistocene time.

Dune origin is rejected because of the long, narrow, smoothly-curving shape of the ridges and the poor sorting of their sands. The absence of closely parallel and branching ridges makes it unlikely that the arcs are beach ridges.

St. George Island (a barrier island) and St. Joseph Spit were selected for additional study. St. George Island consists of two well-defined arcs, both of which are convex gulfward; the intersection of the two arcs is directly opposite the mouth of the bay into which the distributaries of the Apalachicola (that is, Chattahoochee) River enter. The island is about 25 miles long, generally 900 to 1500 feet wide, and has about 25 feet of relief.

St. Joseph Spit is about 13 miles long, between 1500 and 4700 feet wide, and up to 50 feet high (sand dunes). It is convex gulfward.

Seventy-one samples, each weighing about 450 grams, were collected in the field. Each was a spot sample two inches deep and three inches in diameter. Nineteen samples were taken from four traverses across the south sand body, 20 samples in four traverses across the north sand body, 12 samples in four traverses across St. George Island, and 20 samples in four traverses across St. Joseph spit. Organic content (chiefly twigs, leaves, and minute particles of carbonaceous matter) amounted to not more than 3.9 percent by weight in any one sample. Each sample was split to 100 gm and sieved

TABLE 2.—*Decile deviations for 19 samples taken in a grid across north sand body*

	Northern edge			Southern edge	
Eastern end	1.55	1.15	—	1.05	1.04
	1.12	1.22	1.04	1.02	1.05
	1.35	1.26	1.06	1.02	0.88
Western end	0.97	0.88	0.92	0.92	0.73
Average	1.25	1.13	1.01	1.00	0.92

TABLE 3.—*Decile deviations for 20 samples taken in a grid across south sand body*

	Northern edge			Southern edge	
Eastern end	0.81	0.83	0.83	0.75	0.78
	1.08	1.00	1.03	0.98	0.94
	0.85	0.78	0.78	0.72	0.60
Western end	1.26	1.07	1.29	1.03	1.12
Average	1.00	0.92	0.98	0.87	0.86

mechanically for 10 minutes. Cumulated percentages were plotted on probability paper. The results are summarized in table 1. It is concluded that the sands of the north sand body are finer than those of the other three features studied.

Sorting, as measured by phi decile deviation, is better and more uniform on St. Joseph Spit ($\bar{x} = .36$, $\sigma = .020$) than on St. George Island ($\bar{x} = .45$, $\sigma = .12$). Sorting of the sands (of the same size) on the south arc is relatively much poorer. These data may be restated as follows: The sands of neither arc are nearly as well sorted as the sands of the barrier island and spit studied. The sands of the south arc, however, are roughly the same size as those of the island and the spit, whereas the sands of the north arc are finer.

Four sample traverses were made across each arc (giving a total of eight traverses). The phi decile deviation, $(D_9 - D_1)/2.6$, for each sample is shown in tables 2 and 3. Each row of figures represents a single traverse, in which the first figure given (left end) is the northern edge, and the last figure (right end) is the southern edge. The averages are such that it appears that the sands on the southern edge of the north sand body are better sorted than those on the northern edge.

Most of the possible modes of origin were rejected for various reasons as stated earlier. By this method of elimination it is concluded that the two sand bodies are abandoned barrier islands or bars. A comparison with a modern barrier and a modern spit suggest that the two sand bodies did not develop to the extent that the modern features have. This may have been because they were exposed to wave action for a shorter period of time, or because the coastal energy level was lower at that time than it is along the coast now, or for some other reason.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 345–350
FIGS. 1–4, SEPTEMBER, 1958

ELECTRON MICROGRAPHY OF TILL MATRIX[1]

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ABSTRACT

Wisconsin tills from northeastern Ohio were studied under the electron microscope as fragments from less than 1 to several microns and on surfaces as large as $\frac{1}{4}$ inch, using carbon replica techniques. Subhedral "grains" are composed of illite and chlorite; anhedral "grains" are kaolinite. Kaolinite is absent from "late" Cary (just pre-Defiance) till. Nonclay minerals, quartz, feldspar, and calcite, are either absent or exceedingly rare in less than 2 micron sizes, but increase in amount in larger sizes and are common in larger than 4 micron grain sizes. The platy clay minerals are oriented more or less parallel to the horizontal plane of the till in place.

INTRODUCTION

For the past few years several workers have been studying the texture, structure, and composition of the glacial deposits in northeastern Ohio and northwestern Pennsylvania with the support of a National Science Foundation grant to G. W. White. The grain size distribution of these tills has been reported by White and Shepps (1952), Shepps (1953), and Tharin (1958); and Sitler and Chapman (1955) have studied their microfabrics. Structural notations have been made in the field in over 600 outcrops as these exposures were described and sampled. The clay mineral composition of fresh and weathered tills has been determined by X-ray diffraction methods (Droste, 1956a, 1956b; Droste and Doehler, 1957; and Droste and Tharin, 1958), and their heavy mineral content has been investigated by Sitler (1957). The present electron micrographic investigation was undertaken (1) to determine the physical characteristics (roundness, shape, surface texture, size, etc.) of the clay sized grains, (2) to see if the mineral content of the clay sized material already determined by X-ray diffraction could be identified by physical characteristics of the grains, and (3) to see if the schistosity previously reported (Sitler and Chapman, 1955) could be observed.

SAMPLING

Wisconsin tills of several ages have been identified in northeastern Ohio and their surface distribution mapped through a cooperative project of the Ohio Division of Water and the United States Geological Survey already in part reported (White, 1953, pl. 26).

For this study we selected till of Tazewell age near Akron from the excavation of the Universal Clay Products Company, 1 mile west of the center of Mogadore, Springfield Township, Summit County, Ohio (for location see Droste, 1956, fig. 1, No. 1): till from the buried Kent moraine (White, 1957, p. 1902) of "early" Cary age in an excavation for the Ohio Turnpike interchange with Route 14, Streetsboro Township, Portage County, Ohio (Droste, 1956, fig. 1, No. 3, and p. 913): and clay-rich till, immediately pre-Defiance, of "late" Cary age, overlying the coarser Kent till at the Streetsboro location. All the till samples were collected from Horizon V, the blue-gray, unoxidized, unleached, fresh till.

LABORATORY PREPARATION

The clay samples were studied with use of an RCA-EMU-2e electron microscope. Two types of till fragments were used in the investigation.

Till fragments of the first type are prepared in the following way: A clean glass slide is coated with a thin collodion film and very fine particles of till are dusted onto it. The particles are patted gently so that they adhere to the film and the slide is then tapped gently to knock off the larger particles. The slide is then put into a carbon evaporating unit and a very thin layer of carbon is evaporated onto the sample. After the sample has been shadow cast it is scored

[1] Manuscript received March 6, 1958.

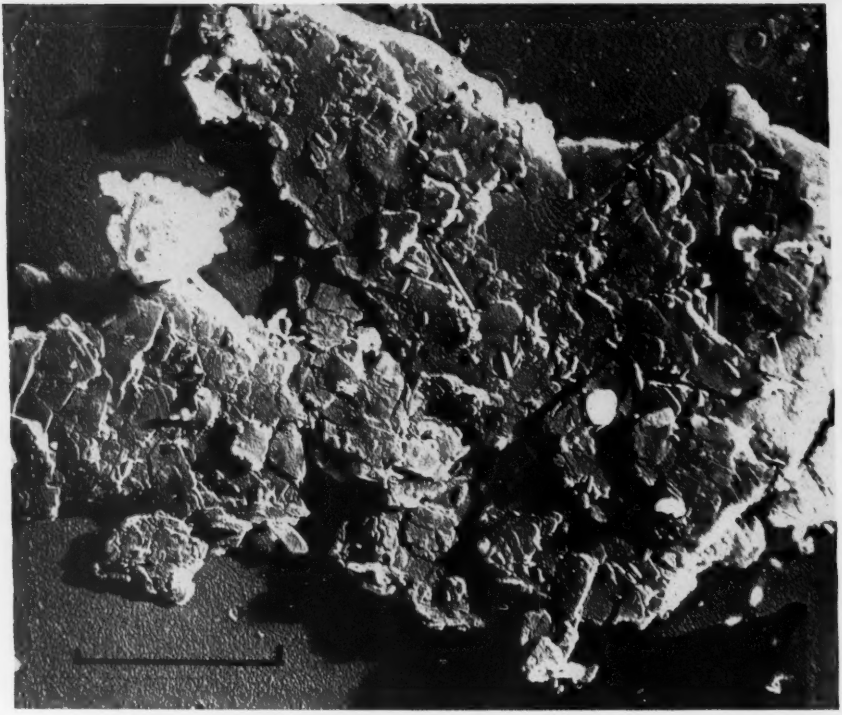

FIG. 1.—Illite and chlorite from Defiance till from Streetsboro, Ohio. The scale indicates one micron.

into grids or squares about ¼ by ¼ in. The slide is slowly put into a dish of clear water and the squares float away from the slide surface. The floating grids are transferred to a dish containing tetrahydrofuran or ethyl alcohol to dissolve the thin collodion film. The washed carbon films are then transferred with an eyedropper to clear water, upon which they float. The floating fragments are transferred by a wire screen to hydrofluoric acid to dissolve the clay from the carbon replica. After about ten minutes the carbon films are observed in the electron microscope to see how well the clay has been dissolved. Many of the carbon films will be inverted when they are initially put into the hydrofluoric acid and the clay will be on the wrong side of the carbon film so that the acid will not attack the clay. These fragments must be discarded. The carbon replica is then washed in water to remove the hydrofluoric acid. The film is then placed on a grid and examined in the electron microscope.

For the second type of examination much larger fragments of the till (up to ½ inch in maximum diameter) are put directly into the carbon shadowing apparatus. Almost all of the surface of the fragments is then coated with carbon. A great deal of difficulty is encountered when the fragment is put into the hydrofluoric acid to remove the clay. Much time is required for the acid to dissolve the silicate minerals. In fact, very few carbon replicas are obtained from which all of the clay material is removed. The first method is simpler and one which produces more clay-free films, but the second must be used in orientation studies.

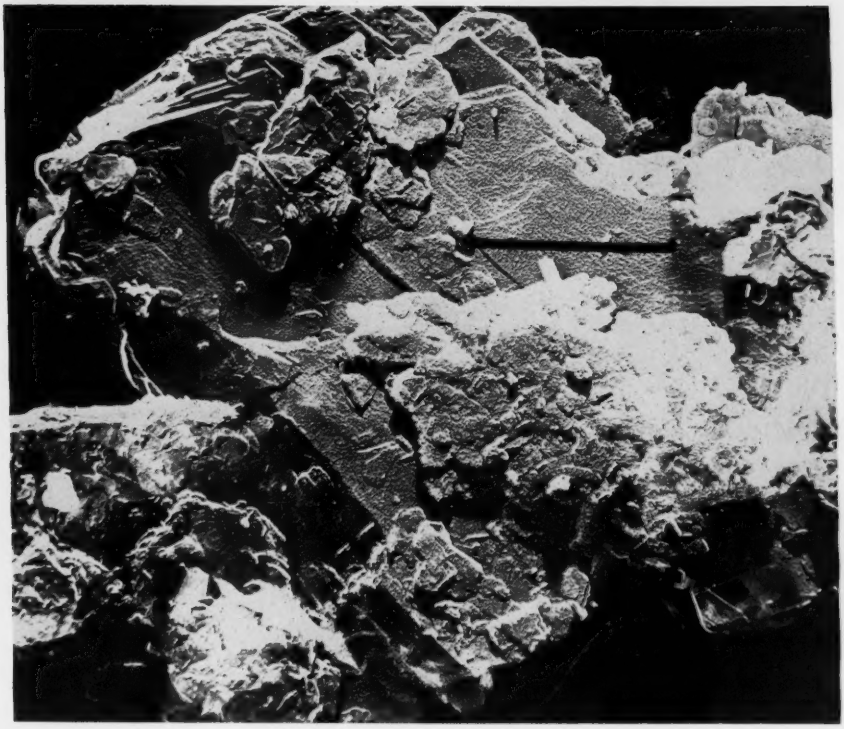

FIG. 2.—A large grain of kaolinite covered in the lower right center by illite and chlorite. Tazewell till from Mogadore, Ohio. The scale indicates one micron.

RESULTS

Under the electron microscope the fragments smaller than 4 microns are almost entirely irregular "masses" of subhedral and anhedral grains, as seen in figure 1. These grains appear in a great variety of shapes and are made up of many smaller crystals. The separate grains range in size from less than 0.1 micron to over 4 microns but the dominant size is less than 1 micron. In samples of Tazewell till another kind of grain, usually between 1 and 3 microns, composed of subhedral crystals, figure 2, stands out in strong contrast to the irregular masses illustrated in figure 1. In figure 2 a smaller mass of the irregular anhedral-type grains can be seen plastered to the large subhedral grains.

The most common grains (fig. 1) are intimate mixtures of illite and chlorite as determined by electron diffraction and confirming earlier X-ray diffraction determinations (*Droste, 1956a; Droste and Doehler, 1957*). It is important to point out that this mixture is a physical one only and is not a mixed layer condition. The second kind of grain, of subhedral crystals (fig. 2), found in Tazewell till and occasionally in Kent till, is kaolinite, identified by the characteristic hexagonal flakes and by electron diffraction, confirming earlier X-ray diffraction determinations. It may be noted that identification by X-ray diffraction of minor amounts of kaolinite in the presence of larger amounts of chlorite is often difficult, and that the electron microscope is better able than the X-ray diffractometer to detect small amounts of kaolinite in this assemblage.

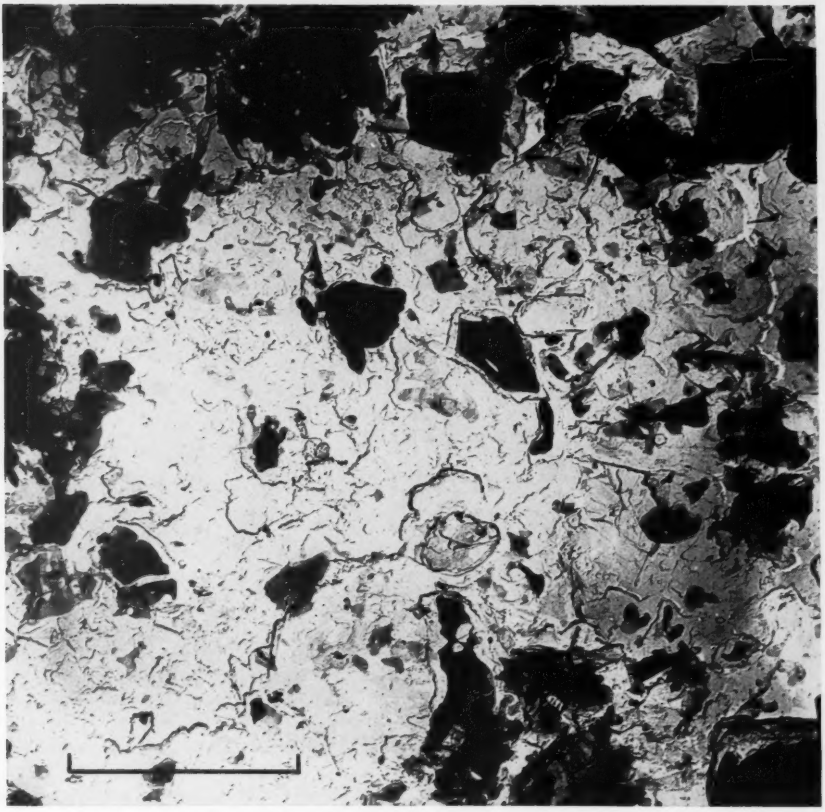

FIG. 3.—Kent till from Streetsboro, Ohio. Surface approximately parallel to the horizontal plane of till in place. Note apparent orientation of basal planes parallel to plane of paper and *C* axes normal to that plane. Dark spots are clay that was not completely removed from the carbon film. A few grains (best seen in the center of the picture) show a weak hexagonal outline and are probably kaolinite. Illite and chlorite are the dominant clay minerals present. The scale indicates one micron.

X-ray diffraction studies (Droste, 1956a) show that chlorite is somewhat more abundant than illite in the 1 to 2 micron size range. Electron diffraction of the irregular masses between 1 and 2 microns indicates that chlorite is more abundant than illite in some of the grains but less in others. All the irregular grains seem to have the same appearance independent of the illite-chlorite ratio. Determination of variation in chlorite-illite content with grain size from electron diffraction data may be possible in further studies.

X-ray diffraction studies show quartz, calcite, and feldspar in small amounts in the less than 2 micron fraction (Droste, 1956a). We hoped the electron micrographic study would show the character of these nonclay minerals. In examination of 150 specimens we saw in the less than 2 micron size fragments only 3 or 4 grains of quartz and no crystals or cleaved fragments of calcite or feldspar. These nonclay minerals are readily observed in the larger than 4 micron sizes, become less apparent in the 4 to 2 micron sizes, and are either absent or very rare in

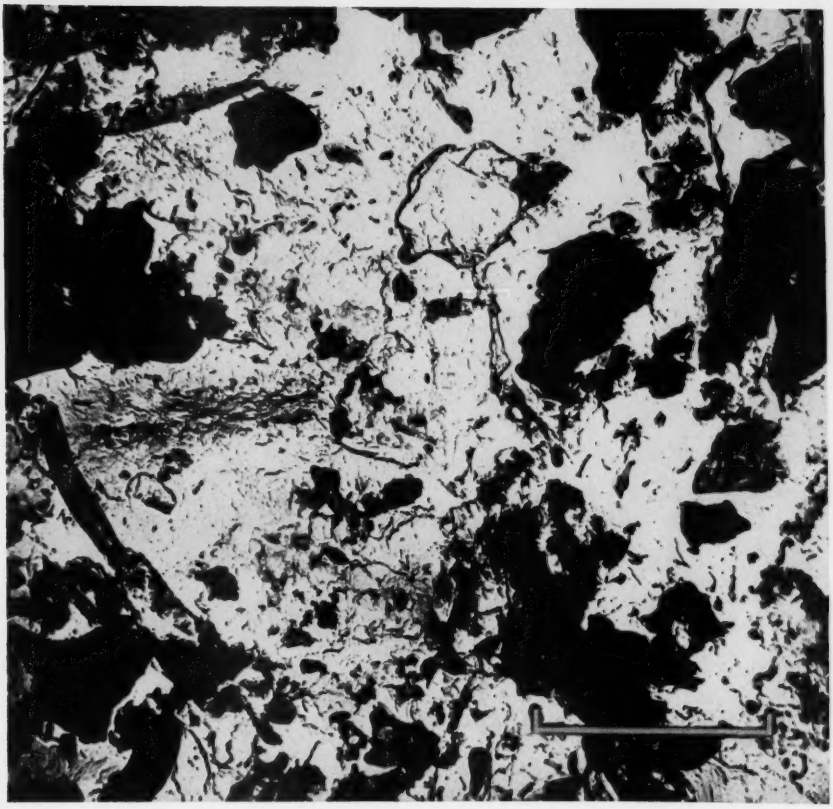

FIG. 4.—Defiance till from Streetsboro, Ohio. Surface approximately parallel to the horizontal plane of till in place. Note apparent orientation of basal planes parallel to plane of paper and *C* axes normal to that plane. Dark spots are clay that was not completely removed from the carbon film. No kaolinite was found in this sample. The scale indicates one micron.

sizes smaller than 2 microns. The presence of quartz, calcite, and feldspar lines in the X-ray diffraction patterns of the less than 2 micron material separated by sedimentation in water (Droste, 1956a) is not compatible with the information obtained from the electron microscope study. This discrepancy is because Stokes' law may not rigidly apply to the smallest grain sizes in fractionation by sedimentation in water. The X-ray lines for quartz, calcite, and feldspar in the less than 2 micron material must, therefore, be almost entirely the result of contamination by grains larger than 2 microns.

From oriented thin section studies, Sitler and Chapman (1955) report that schistosity exists in these tills in a plane roughly parallel to the horizontal surface, and it was hoped that schistosity would be seen in the electron microscope study. A number of attempts were made to obtain carbon replicas from oriented small blocks. It is difficult to maintain orientation when dealing with fragments of till less than $\frac{1}{4}$ inch in diameter and it is also difficult to maintain orientation of a carbon replica after removing it from a till particle. The hydrofluoric acid removes the silicate minerals from the replicas

of block surfaces very slowly and does not remove them completely. The carbon surface therefore is not completely free of foreign material and it is impossible to get an unobstructed look at a clean replica surface. Figures 3 and 4 are replicas of horizontal surfaces. The dark areas are clay flakes that have not been removed by the hydrofluoric acid etching. It certainly appears that in both of these figures the clay minerals are oriented with their basal planes essentially in the plane of the paper and their *C* axes are essentially perpendicular to the plane of the page.

The observations made on the fragments of unknown orientation (figs. 1 and 2) also indicate that some preferred orientation exists in the clay minerals of the till matrix. Although this orientation is not perfectly developed and there is no indication in which plane the orientation is to be found, the orientation definitely does exist. Almost all of the less than 4 micron particles are made up of many smaller mineral flakes all oriented with respect to their basal planes and thus there is a preferred orientation throughout the till matrix. The most logical place for this orientation to exist is in the plane of schistosity. It is hoped with improved techniques of preparation that oriented carbon replicas can be made with greater ease and certainty and direct information obtained concerning the orientation of clay material in the till matrix.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 351–360
FIGS. 1–9, SEPTEMBER, 1958

ROUNDNESS AS AN INDICATOR OF ENVIRONMENT ALONG THE COAST OF PANHANDLE FLORIDA[1]

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ABSTRACT

Fifty-nine sand samples have been analyzed to determine the relationship between depositional environment and quartz grain roundness. Roundness as used in this study is based on the sharpness of the corners and edges of individual sand grains from three size fractions. The environments from which samples were taken are: off shore, dune, beach ridge, open swash zone, protected swash zone, marsh, lagoon, and river. Sands from certain groups of environments show distinctive roundness patterns.

INTRODUCTION

The purpose of this study is to find if there is a relationship between depositional environment and quartz grain roundness in certain recent coastal sediments of Panhandle Florida.

The area studied is in Gulf and Franklin Counties along the southern coast of Panhandle Florida. Specific areas of sampling were St. George and St. Vincent islands, which lie in the southwestern part of Franklin County; Indian Peninsula and St. Joseph's Spit, which lie in the southern part of Gulf County; and bays and mainland shores of both counties. Field work was done in the summer of 1956.

MacCarthy (1935) investigated the size and shape relations of eolian (dune and desert) and beach sands. He concluded that eolian sands have a greater degree of roundness than beach sands and that this is even more evident when the same size-fractions of each sample are compared. He reported this difference in roundness in all grain sizes of eolian and beach sands. When MacCarthy compared beach dune and desert sands, the beach dunes had a higher degree of roundness not only when comparing size-fraction for size-fraction but also as whole samples.

Roundness studies have been made along the Texas and Louisiana coasts by Beal and Shepard (1956) in an effort to determine the relationship of roundness values to environments of deposition. After studying beach and dune sands they concluded that there is a measurable and consistent difference between beach and dune grain roundness.

PHYSIOGRAPHY

The area under study is characterized by loose sand, low relief, and poor drainage.

The area consists of bays, coves, lagoons, marshes, swamps, barrier islands, a spit, and a peninsula. The barrier islands, peninsula, and mainland are diversified physiographically. There are extensive dune tracts, some extending as much as a quarter of a mile from the beach inside the long sand beaches on the Gulf side. Inland beyond the beaches on St. Vincent Island and at the east and west ends of St. George Island are clusters of beach ridges, aligned parallel with the shore, in belts up to three miles wide. Beach ridges extend the full width of St. Joseph's Spit and lie parallel to the shore. On the mainland there are sparse dune tracts along the beaches. Beach ridges west of Apalachicola trend east and west parallel to the bay shore in a belt two to three miles wide. In the Port St. Joe area there is another cluster of beach ridges aligned north and south, parallel with the bay shore, extending inland two to three miles.

Marsh environments are the most extensive on the mainland. Sand, high in carbonaceous material, was found in marsh and lagoon environments and in the troughs between the beach ridges.

[1] Based on a thesis submitted to the Faculty of Florida State University in partial fulfillment of the requirements for the degree, Master of Science. The writer is grateful to W. F. Tanner for direction of the study and to B. F. Buie and S. S. Winters for helpful suggestions. Illustrations were drafted with the assistance of H. J. McCoy. Manuscript received March 14, 1958.

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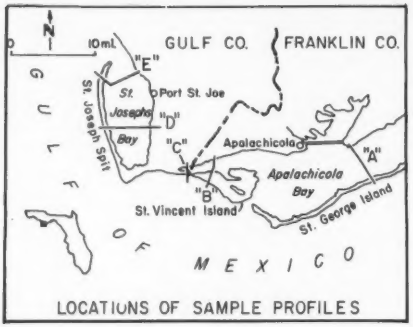

FIG. 1.—Locations of sample profiles.

The vegetation consists of a sparse palm and pine tree cover with a sparse to thick grass cover on the dunes and beach ridges and in marsh environments. In the field there is a marked distinction between the vegetation cover on the crests and in the troughs of the beach ridges.

COLLECTION OF SAMPLES

The collection of field samples was patterned after the procedure outlined by Beal and Shepard (1956).

Samples for roundness measurement were collected along predetermined profiles (fig. 1) from the following environments, where present: off shore, open swash zone, protected swash zone, dune, beach ridge, marsh, lagoon, and river. River samples were collected by Haley (1956). Care was taken not to collect samples from certain disturbed areas.

A coffee can, cut to a depth of two inches, was used to obtain a sample from the top two inches of the sand deposit in each environment. Diving operations were conducted from a boat in the collection of off shore and lagoon samples.

MECHANICAL ANALYSIS

Five samples from each environment were selected for use in the mechanical analysis. Samples containing organic material were washed and the organic material decanted. All samples were then spread out to dry at room temperature. After drying, each sample was divided in a riffle sample splitter to obtain a representative portion of 100 grams.

Mechanical separation was performed on a Tyler Ro-Tap machine with a 10-minute operating time. A complete set of fourth-root-of-two Tyler screens was used in this separation.

Cumulative-frequency curves were plotted on logarithmic probability paper, plotting percent against diameter. Six of these are presented in figures 2 and 3. The sorting coefficient (S_0), the percentile deviation

$$\left(=\frac{P_{97.5}-P_{2.5}}{4}\right),$$

and the median diameter are presented in table 1.

The grain size distributions for 31 of the samples are clearly unimodal and for nine perhaps bimodal or polymodal. The bimodal or polymodal distributions represent water environments.

Although averages of some sedimentary parameters show slight differences between environments, specific parameters show no consistent difference.

ROUNDNESS ANALYSIS

Roundness has been a confusing term, widely misused to refer to sphericity or shape. Roundness was defined by Wadell (1932, 1933, 1935) as the ratio of the average radius of the corners and edges to the radius of curvature of the maximum inscribed sphere when projected to a standard diameter of 70 mm. Roundness is simply related to the sharpness of the corners and edges of the grain particle and should be distinguished from sphericity. Pettijohn (1949, p. 53) suggested that roundness is an index to the maturity of a sediment and distance of transportation.

Wadell defined sphericity as the diameter of a sphere of the same volume as the particle divided by the diameter of the circumscribing sphere. Pettijohn suggested that the final sphericity of a sand grain appears to be determined by its original shape. He also stated that roundness and sphericity are both functions of grain size and, therefore, are functions of each other. He illustrated the relationship between roundness and sphericity by showing that sand grains with the same value of sphericity may have any degree of roundness. Further, sand grains with the same roundness may differ in sphericity.

If roundness, as defined above and distin

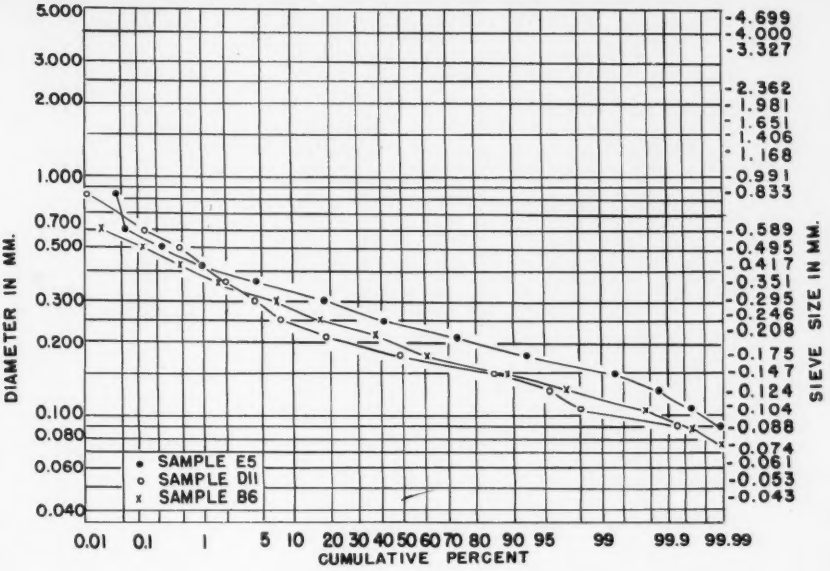

FIG. 2.—Beach ridge environment

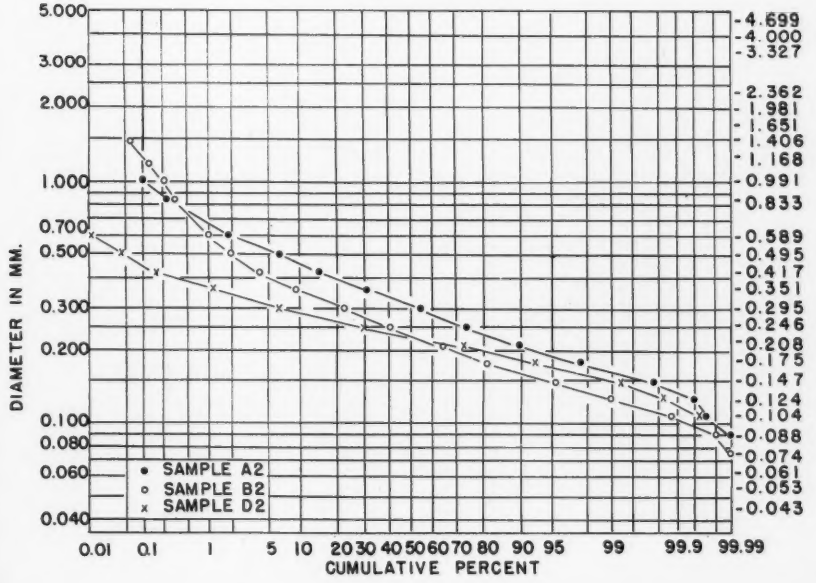

FIG. 3—Open swash zone environment.

TABLE 1.—*Sedimentary parameters*

	Environment	Median mm	S_0	Median ϕ	$\frac{P_{97.5}-P_{2.5}}{4}$
Max.		.308	1.38	2.11	.622
Avg.	Off shore	.268	1.23	1.90	.467
Min.		.231	1.15	1.70	.267
Max.		.305	1.25	2.18	.490
Avg.	Open swash zone	.241	1.21	2.05	.401
Min.		.220	1.11	1.71	.277
Max.		.288	1.20	2.46	.610
Avg.	Dune	.226	1.17	2.15	.399
Min.		.182	1.14	1.80	.302
Max.		.296	1.20	2.54	.422
Avg.	Beach ridge	.221	1.17	2.18	.364
Min.		.172	1.14	1.76	.320
Max.		.310	1.27	2.51	.562
Avg.	Marsh	.233	1.22	2.10	.470
Min.		.176	1.13	1.69	.402
Max.		.435	1.22	2.56	.592
Avg.	Protected swash zone	.271	1.20	1.88	.480
Min.		.170	1.16	1.20	.402
Max.		.278	1.28	2.46	.532
Avg.	Lagoon	.236	1.22	2.08	.432
Min.		.182	1.17	1.85	.340
Max.		.760	1.58	2.11	.757
Avg.	River	.420	1.42	1.25	.679
Min.		.232	1.25	.40	.577

guished from sphericity, is related to the maturity of a sediment and distance transported, then possibly there is a useful relationship between roundness and depositional environments.

The roundness study was patterned after Beal and Shepard (1956) except for one modification. Beal and Shepard studied the composite quartz grain roundness between 0.124 mm and 0.061 mm. The present study was made on grains between 0.208 mm and 0.175 mm, between 0.104 mm and 0.088 mm, and, where available, between 0.053 mm and 0.043 mm. This modification was made in order to discover any significant differences in roundness between size fractions in various depositional environments.

To avoid bias in roundness determination, the writer arranged to have each sample fraction renumbered by another person.

Sieve fractions were put through a Sepor microsplitter to obtain a few hundred representative grains. The grains were brushed carefully onto a black counting tray, which was tapped to orient the grains on their flattest faces. Parallel traverses were made across the tray. All quartz grains coming in the field of view of the binocular microscope were counted until 50 to 100 grains had been counted.

Each grain was placed in one of Powers' (1953) class intervals based on the criteria in table 2. These criteria were modified from Pettijohn (1949) to fit Powers' six-fold classification. The roundness averages are presented in table 3.

The majority of the 59 samples came from dunes and beach ridges. A minimum of five samples were studied from the other six environments. All roundness values for each size fraction were plotted against environment (figs. 4, 5, and 6). A line connecting the average roundness of each environment is presented.

TABLE 2.—*Description of roundness grades*

Grade terms	Criteria
Very angular	Showing *no* evidence of wear; corners sharp. Numerous secondary sharp corners
Angular	Showing very little evicence of wear on corners; no wear on edges. Several secondary sharp corners
Subangular	Showing definite effects of wear. Grains still have original form and faces are virtually untouched; the edges have been rounded off to some extent. Some secondary corners
Subrounded	Showing considerable wear. The edges are rounded off to smooth curves, and the areas of the original faces are considerably reduced but the original shape of the grain is still distinct. Very few secondary corners; much rounded
Rounded	Original faces almost completely destroyed, but some comparatively flat surfaces may be present. All original edges and corners have been smoothed off to rather broad curves. Secondary corners greatly subdued
Well rounded	No original faces, no edges or corners left. The entire surface consists of broad curves; flat areas are absent. No secondary corners

Figure 4 shows a plot of roundness for grains between 0.208 and 0.175 mm in diameter. Beach ridge and lagoon sand grain roundness is consistently greater than that of other environments, with the marsh and protected swash zone generally lower. A plot of roundness for grains between 0.104 and 0.088 mm (fig. 5) shows three distinguishable groups. The beach ridge and open swash zone have a greater degree of roundness. The marsh and protected swash zone grain roundness is generally lower with all other environments in an intermediate group. Figure 6 (roundness for grains between 0.053 and 0.043 mm) gives another pattern with the roundness in the beach ridge and river higher and marsh lower than the other environments.

A comparison of these size fractions (figs. 4, 5, and 6) shows that the finer size fraction has a lower degree of roundness than the coarser sizes.

The average roundness of various environments investigated by Beal and Shep-

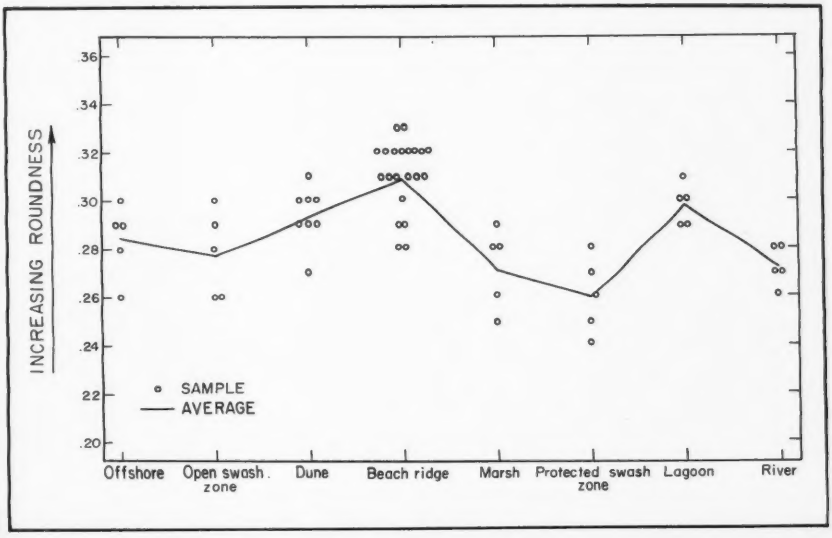

FIG. 4.—Roundness values for grains between 0.208 mm. and 0.175 mm.

TABLE 3.—*Roundness values*

Environment	Number	0.208–0.175 mm	0.104–0.088 mm	0.053–0.043 mm	Environment average
Off shore	A1	.26	.29	.22	
	B1	.28	.28	.21	
	C1	.30	.29	.23	
	D1	.29	.27		
	E1	.29	.25		
	Avg.	.284	.276	.220	.260
Open swash zone	A2	.29	.30		
	B2	.26	.30		
	C2	.26	.28	.21	
	D2	.28	.35		
	E2	.30	.35	.21	
	Avg.	.278	.316	.210	.268
Dune	A3	.30	.26		
	A4	.31	.27		
	B3	.29	.29		
	B4	.29	.30	.24	
	C3	.30	.29	.23	
	C4	.30	.30	.23	
	C5	.29	.30	.23	
	E15	.27	.28	.22	
	Avg.	.293	.286	.230	.269
Beach ridge	B5	.32	.32	.24	
	B6	.28	.29	.23	
	C9	.32	.32	.26	
	C10	.31	.31		
	C11	.31	.31		
	C12	.31	.33		
	D3	.31	.32		
	D4	.32	.34		
	D5	.32	.29		
	D7	.32	.30	.25	
	D11	.31	.31		
	D12	.31	.28		
	E3	.29			
	E4	.32	.30		
	E5	.33	.32		
Beach ridge	E6	.32	.32		
	E7	.29	.31		
	E8	.33	.29		
	E10	.28			
	E11	.32	.29		
	E12	.30	.29		
	Avg.	.310	.307	.245	.287
Marsh	B7	.26	.25	.21	
	C8	.25	.27	.21	
	D10	.28	.25	.19	
	D6	.28	.25		
	E9	.29	.25		
	Avg.	.272	.254	.203	.243
Prot. swash zone	A6	.26	.26	.23	
	B9	.24	.26	.22	
	C6	.27	.27	.23	
	D9	.28	.27	.23	
	E14	.25	.25		
	Avg.	.260	.262	.227	.249

TABLE 3.—*continued*

Environment	Number	0.208–0.175 mm	0.104–0.088 mm	0.053–0.043 mm	Environment average
Lagoon	A5	.30	.31	.22	
	B8	.29	.28	.23	
	C7	.31	.27	.21	
	D8	.29	.30	.21	
	E13	.30	.30		
	Avg.	.298	.292	.217	.269
River[1]	H3	.26	.28	.25	
	H7	.27	.28	.28	
	H16	.28	.28	.24	
	H23	.27	.29	.28	
	H30	.28	.30	.25	
	Avg.	.272	.286	.260	.272

[1] Collected by Haley (1956).

ard (1956) is compared in figure 5 with the roundness values of the size fraction between 0.104 and 0.088 mm. Further, the average of all the roundness values in this study is compared with the work of Beal and Shepard in figure 7. There is less roundness in the sediments from the various environments in this study than in those in Texas and Louisiana.

A difference in roundness for each size fraction was found between environments along the profiles sampled. This is shown in

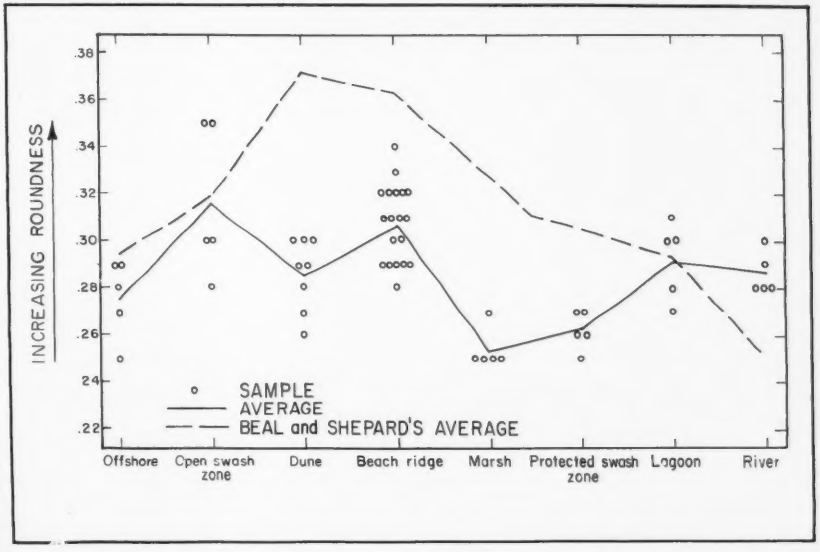

FIG. 5. Roundness values for grains between 0.104 mm. and 0.088 mm. The average roundness of various environments investigated by Beal and Shepard is compared.

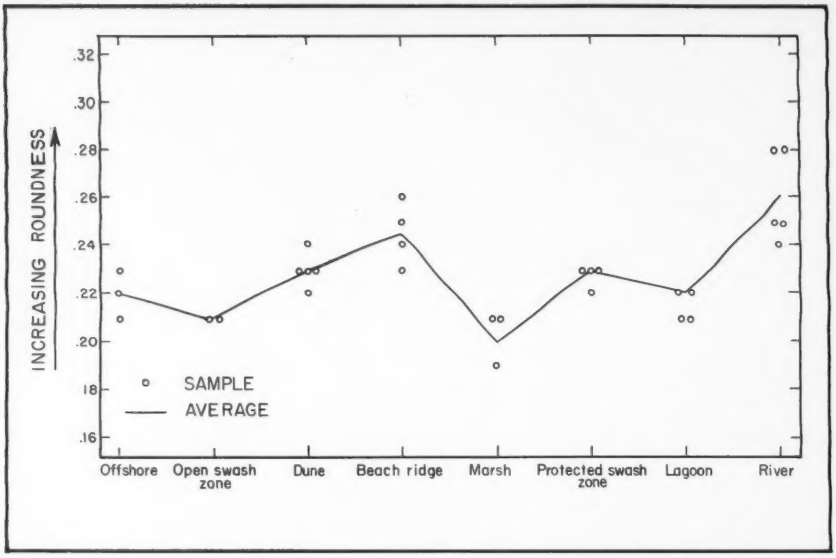

FIG. 6.—Roundness values for grains between 0.053 mm. and 0.043 mm.

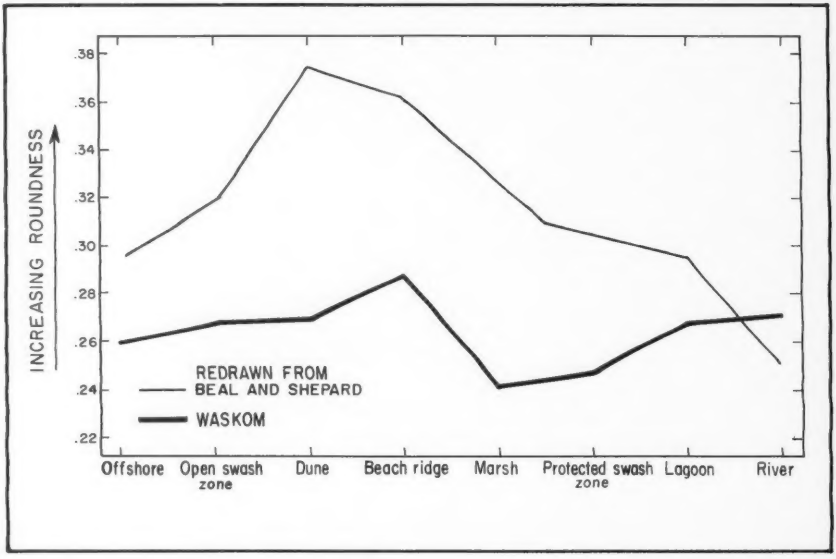

FIG. 7.—A comparison of the average of all the roundness values with those of Beal and Shepard.

figure 8. The difference in roundness between the crest and trough, or base of beach ridges and sand dunes, as shown in individual profiles, is negligible.

Figure 9 shows the high and low limits of roundness for each size fraction in the respective environments.

SUMMARY AND CONCLUSIONS

The grain size distributions in all environments are generally unimodal. Deviation from unimodal distribution occurs in environments where the sediments were deposited by water. The median diameter and sorting coefficient when plotted against one another show no clear correlation with environments. Other sedimentary parameters examined in the sedimentary analysis also indicate no recognizable difference between depositional environments.

Sands from certain groups of environments show distinctive roundness patterns. In all size fractions beach ridge environments have a high degree of roundness and marsh environments have a low degree of roundness. The protected swash zone environments have a low degree of roundness in the coarser size fractions and have intermediate roundness in the finer size fractions. All other environments in all size fractions range from a high degree of roundness to a low degree of roundness. A noticeable difference exists between the roundness of the two coarser size fractions and the finer size fractions in that the latter has a lower degree of roundness than the two coarser sizes. Pettijohn (1949, p. 53) found a similar relationship.

A comparison of quartz grain roundness with the data published by Beal and Shepard shows there are similarities between the thesis area and the Texas and Louisiana coast (fig. 9). However, the results in the thesis area do not show differences between environments as clearly cut as those reported by Beal and Shepard.

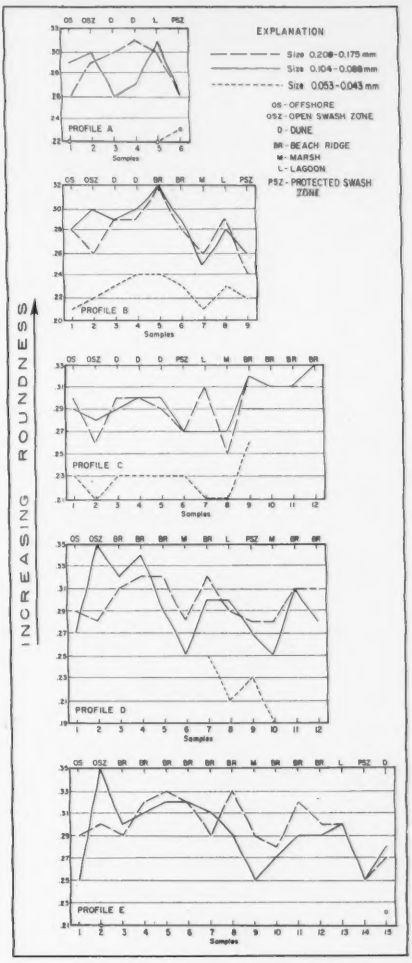

FIG. 8.—Roundness relationship of surface samples for each size fraction along each profile.

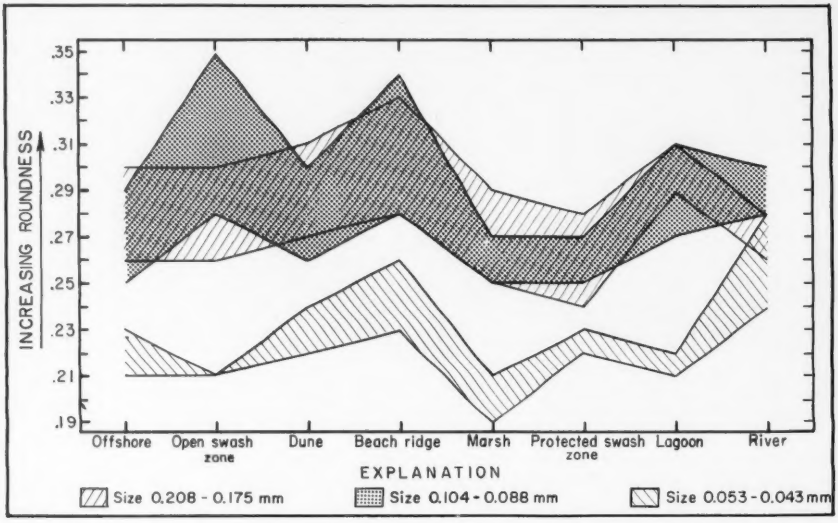

FIG. 9.—High and low limits of roundness for each size fraction in the respective environments.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 361–365
SEPTEMBER, 1958

SIZE DISTRIBUTION OF ZIRCON AND TOURMALINE GRAINS IN SOME SAMPLES OF THE LISSIE FORMATION[1]

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ABSTRACT

Zircon and tourmaline grains in samples of the Lissie Formation near Houston, Texas, appear to be log-normally distributed by number percentage. Size distributions of bulk samples (predominantly quartz) are log-normal by weight percentage. The similarity of the distributions may be caused by hydraulic properties of the transporting agent. The size distribution of the heavy minerals may, however, be related solely to distributions in primary source rocks, and the distribution of the light minerals may be controlled entirely by fracturing during the sedimentation process.

INTRODUCTION

The Lissie Formation (Deussen, 1914) is a sequence of poorly indurated Lower Pleistocene sediments. The formation crops out along the Gulf of Mexico in Texas in a broad band roughly parallel to the present coast and 50 to 100 miles inland. The sandy sediments of this formation contain a small amount of heavy minerals, of which zircon and tourmaline are the most abundant. The present study is largely concerned with the size distribution of these two minerals.

A total of 28 samples was taken along the strike of the Lissie Formation from the Colorado River, on the west almost to the Trinity River on the east. The samples represent the upper sandy member of the formation as defined by Deussen (1914), which is equivalent to the Lissie Formation in the terminology of Doering (1935, 1956). Mineralogically, the portions of the formation studied consist almost entirely of quartzose sand and silt with very minor amounts of heavy minerals and some clayey material.

The samples studied were analyzed by methods described in the following section. Complete data on the various samples are given in a Master's thesis by one of the authors (Dawson) written in 1958 and entitled "A Study of the Non-Opaque Heavy Minerals in the Lissie Sandstone," which is on file at The Rice Institute. One of the authors (Rogers) takes complete responsibility for the statistical analysis of the data in the present paper.

[1] Manuscript received March 22, 1958.
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The work reported in this paper was supported by the Harry Carothers Wiess fund of the Department of Geology of The Rice Institute. The writers thank Dr. Carey Croneis for suggesting the present study.

PROCEDURE

The procedures used to obtain the size distributions and abundances of heavy minerals in samples of the Lissie Formation are as follows:

1.—All samples were taken several feet back of the outcrop surfaces and were poorly cemented and highly disaggregated.

2.—About 10 to 15 grams of each of the disaggregated samples were boiled for about 20 minutes in a mixture of 10 grams of stannous chloride and 20 cc of concentrated hydrochloric acid in order to remove iron oxide and calcite. The samples were then washed and dried. The effects of such treatment on heavy minerals are discussed by Krumbein and Pettijohn (1938, p. 355). The maximum weight loss by this procedure was 12.5 percent, and the average for all samples was 6.5 percent.

3.—Four of the samples studied contained too much clay (more than about 10 percent) for effective removal of the heavy minerals. These samples were allowed to stand for several hours in sodium oxalate solution after the treatment with hydrochloric acid-stannous chloride. Following this deflocculation, the samples were again washed and dried.

4.—Heavy minerals were separated from the dried, disaggregated samples from which the iron oxide and calcite had been removed. The samples were placed in bromoform in

celluloid centrifuge tubes and were centrifuged for 10 minutes at 2500 rpm, stirred, and recentrifuged a total of four times. After centrifuging, the bromoform was frozen, and the bottom of the centrifuge tubes (containing the heavy minerals) was cut off. The bromoform was then melted and the heavy minerals were separated, washed in acetone, and mounted in Canada balsam.

5.—Mineral percentages were determined from a count of 300 grains. Size distributions of zircon and tourmaline were determined by measurement of the long axes of 300 grains of each mineral as they appeared in the slide. The accuracy to be expected from grain counts is discussed by Dryden (1931).

6.—Cumulative frequencies of the size distributions were plotted on logarithmic probability paper, and parameters of the distributions were determined from these curves. Sizes were measured and plotted in intervals based on integral powers of $\sqrt{2}$ rather than on integral powers of 2.

The size distributions of the bulk sample were made on material from which the iron oxide and calcite had been removed. The distribution was determined by placing 50 grams of sediment in sieves in a RoTap mechanical shaker for 10 minutes. The weight percentages obtained were plotted on logarithmic probability paper.

Statistical analyses of the size distribution data were made by conventional chi-square methods. Expected frequencies in the measured size intervals were determined from log-normal curves (straight lines on logarithmic probability paper) drawn as the best approximation to the plotted observations.

ANALYTICAL RESULTS

The major non-opaque heavy minerals in the Lissie Sandstone are zircon and tourmaline, with minor amounts of garnet, staurolite, rutile, sphene, kyanite, and epidote. In samples taken from the drainage area of the Colorado River the percentage of hornblende is high (over 60 percent), but in other areas the hornblende comprises less than 5 percent of the non-opaque heavy minerals from each sample. The abundance of hornblende in sediments near the Colorado River has been reported by Cogen (1940) and correlates well with the abundance found in sands of the modern Colorado River by Bullard (1942). With the exception of hornblende, the non-opaque heavy minerals of the Lissie Formation are all stable varieties. Table 1 lists the average relative abundance and range of abundances of the non-opaque heavy minerals in studied samples taken outside of the Colorado River drainage basin.

TABLE 1.—*Relative abundance of heavy minerals in studied samples of the Lissie Formation near Houston, Texas*

Mineral	Average abundance (in number percent)	Range of abundances (in number percent)
Zircon	57	68.7—34.3
Tourmaline	22	49.0—10.0
Rutile	6.5	11.0— 2.0
Epidote	3.8	11.3— 0.0
Staurolite	2.9	7.7— 0.7
Sphene	3.1	12.3— 0.0
Garnet	1.6	6.0— 0.0
Hornblende	1.9	5.3— 0.0
Kyanite	1.2	3.0— 0.0

The averages and ranges of the medians, sorting coefficients, and skewnesses of the size distributions of zircon and tourmaline in the Lissie Formation are shown in table 2. The zircon grains were measured in 21 samples and tourmaline grains in 13 samples; the scarcity of tourmaline in some samples prevented adequate measurement. The distributions of zircon and tourmaline grains are plotted on the basis of the number of grains in each size interval, not on the basis of weight as in the case of the size distribution of the bulk sample.

As shown in table 2, the zircon and tourmaline grains are very well sorted, and their size distributions are essentially unskewed. The restriction of the grains to relatively few size classes prevents accurate measurement of skewness and, especially, of kurtosis. The available size frequencies, however, if plotted as cumulative curves on logarithmic probability paper, yield approximately straight lines, and thus the size distributions appear to fit a log-normal curve. Chi-square tests for the significance of apparent deviations from log-normality were made as previously described. The results are shown in table 3. Except for one tourmaline and two zircon samples the various distributions show no good evidence for deviation from log-normality, and in many samples the chi-square

TABLE 2.—*Median, sorting coefficient, and skewness of the size distributions of zircon and tourmaline in samples of the Lissie Formation near Houston, Texas*

	Zircon	Tourmaline
Median (average)	0.082 mm	0.094 mm
Median (range)	0.100–0.051 mm	0.112–0.064 mm
Sorting coefficient (average)	1.26	1.23
Sorting coefficient (range) $\sqrt{Q_3/Q_1}$	1.38–1.12	1.32–1.17
Skewness (average)	1.00	0.99
Skewness (range) Q_3Q_1/M^2	1.08–0.900	1.02–0.970

value is extremely small or apparently zero. The lack of evidence for deviation does not prove, of course, that the distributions are actually log-normal; some curve other than the normal might also fit the data.

The average and range of the medians, sorting coefficients, and skewnesses of 14 bulk samples (essentially quartz) of the Lissie Formation are shown in table 4. Plots of the cumulative size frequencies (in weight percentages) on logarithmic probability paper show no detectable deviation from log-normality for 11 of the samples; for the other 3 samples the chi-square for deviation from log-normality is small. The size distribution of the bulk sample shows excellent sorting and is essentially unskewed.

TABLE 3.—*Chi-square values for deviation of zircon and tourmaline size distributions from log-normal curves*

Zircon		Tourmaline	
Chi-square	Degrees of freedom	Chi-square	Degrees of freedom
0 (approx.)	2	0 (approx.)	2
0 (approx.)	3	0 (approx.)	3
0 (approx.)	5	0 (approx.)	5
0.41	5	0 (approx.)	5
0.57	4	0 (approx.)	5
1.02	5	0.28	3
1.34	4	1.01	4
1.42	5	1.32	4
1.47	5	2.30	3
1.56	4	2.86	6
1.87	4	3.53	4
2.27	5	5.25	4
2.65	5	14.73	4
2.88	6		
3.25	6		
3.84	4		
4.01	6		
4.70	2		
5.83	4		
6.25	4		
11.66	5		

DISCUSSION

The predominance of quartz and the absence of heavy minerals other than the most stable varieties throughout much of the Lissie Formation indicates that the Lissie sands were derived largely from the erosion of sedimentary source rocks. Presumably, sediments were the only source rocks available in the drainage areas of the Brazos and Trinity Rivers, but the igneous and metamorphic rocks of central Texas in the Colorado River drainage basin contributed some unstable materials (for instance, hornblende) to the Lissie Formation in that area. The sediment derived from these various sources was deposited in a broad coastal plain-deltaic environment along the Early Pleistocene shoreline.

Apparently the heavy minerals (zircon and tourmaline) and the light minerals (measured as the bulk sample, predominantly quartz) exhibit the same types of dis-

TABLE 4.—*Median, sorting, coefficient, and skewness of the size distributions of bulk samples (quartz) of the Lissie Formation*

Median (average)	0.165 mm
Median (range)	0.267–0.105 mm
Sorting coefficient (average)	1.40
Sorting coefficient (range) $\sqrt{Q_3/Q_1}$	1.56–1.29
Skewness (average)	1.00
Skewness (range) Q_3Q_1/M^2	1.12–0.925

tribution curves in the studied samples of the Lissie Formation. The data described previously show that the quartz is log-normally distributed by weight percentage, and the maximum diameters of the heavy minerals may be log-normally distributed by number percentage. These distributions are essentially the same provided that the shapes of the grains of each mineral do not vary with size. (If x = diameter and $\log x$ is normally distributed, then kx^3 = volume, and $\log kx^3 = \log k + 3 \log x$ must also be normally distributed; the mathematics of distribution curves is discussed by Epstein, 1947.) An apparent log-normal distribution by number percentage has been reported previously by Rogers and Powell (1958) for zircon grains in the lower part of the Beaumont Formation. A study of river sands by Rittenhouse (1943) showed that both light and heavy minerals have similar shapes of size distribution curves in the same sample.

The similarity of the size distributions of light and heavy minerals in the Lissie Formation is interesting in view of the following considerations:

1.—The size distribution of each mineral species in primary rocks (igneous and metamorphic) may be log-normal. This conclusion can be justified theoretically by consideration of growth processes in homogeneous media, and some empirical evidence is available. Kottler (1950) reviews previous literature on size distributions and discusses the experimental and theoretical basis for log-normal distributions. Extensive data on the size distributions of silver halide grains in photographic emulsions are given by Wightman, Trivelli, and Sheppard (1924).

2.—Heavy minerals, especially those occurring as accessory minerals in primary rocks, are generally incorporated in sediments without extensive fracturing, commonly as grains which are only slightly abraded from their former sizes. This lack of fracturing is apparent even in grains which have undergone multiple reworking from their original source rocks, as is the case for minerals in the Lissie Formation.

3.—Light minerals from primary source rocks are incorporated in sediments only after extensive fracturing during weathering, erosion, and transportation. The size distribution of freshly weathered material has been shown to follow Rosin's law for crushed material (Krumbein and Tisdel, 1940). Repeatedly and thoroughly crushed material may, however, exhibit a log-normal size distribution (for discussion see Hatch and Choate, 1929). Other types of distribution are also possible (Roller, 1937). In some cases the shapes of distribution curves may depend somewhat on the type of measurement taken to represent "size." So many sediments exhibit essentially log-normal distributions by weight percentage for the bulk sample that apparently the total of the fracturing processes active during weathering, erosion, and transportation tend to produce a log-normal distribution, as in repeatedly crushed material. Actually, the size distribution of material which follows Rosin's law could be readily converted into a log-normal distribution by slight fracturing of the coarse sizes into finer material.

In view of the much greater importance of fracturing in controlling the size distribution of light minerals than of heavy minerals in the Lissie Formation, and possibly in many other sediments, it is interesting that the shapes of the distribution curves of the different minerals are the same. Both heavy and light minerals have the same distribution (except for possible difference in sorting coefficients) as in original source rocks, despite the fact that Lissie sediments have undergone extensive recycling and despite the fact that the light minerals have passed through a stage (freshly weathered material) in which their size distribution differs from log-normality. The similarity of the distributions may be caused by hydraulic conditions in streams, or other currents, which naturally impose log-normal distributions on the contained sediment. The fact that the observed distributions in the deposited sediments, however, can either be produced by repeated fracturing, in the case of light minerals, or can be matched closely by the distribution in source rocks, in the case of heavy minerals, indicates that sorting and other stream processes are not a major influence. Possibly transportational agents have no effect on the shapes of distribution curves, although they certainly influence the medians of the distributions.

It seems likely that a study of distributions, such as the analysis outlined above, may be useful in many geologic investigations. It may be possible to determine which of several influencing factors has been dominant in the formation of some geologic feature. The question as to whether a log-normal distribution represents randomness or some particular controlling force must, of course, enter such an analysis, but a discussion of that problem is beyond the scope of this paper.

NOTE ADDED IN PROOF

The writers' attention has been called to a recent book entitled *The Lognormal Distribution*, Monograph 5 of the University of Cambridge Department of Applied Economics, by J. Aitchison and J. A. C. Brown, published in 1957 by the Cambridge University Press. This book discusses many theoretical and applied aspects of the subject. The relation between weight and number frequency distributions of particle sizes is described on pages 100–101.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 366–371
FIGS. 1–2, SEPTEMBER, 1958

RATE OF FORMATION OF A DIAGENETIC CALCAREOUS CONCRETION[1]

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ABSTRACT

Age determinations by the C^{14} method have been made on the matrix and enclosed fauna of a Quaternary calcareous concretion from the shelf southeast of Cape Campbell, New Zealand. The results show that the concretion must have formed within 19,500 years, and probably within 7500 years. It appears to have grown in a calcareo-argillaceous sediment containing an abundant molluscan fauna.

The classification of concretions, and the terminology applied to them, are discussed in this paper. The three categories recognised by Richardson (1921) are retained, but it is suggested that the terms "contemporaneous," "penecontemporaneous," and "subsequent" should be replaced by "syngenetic," "diagenetic," and "epigenetic." The Cape Campbell concretion is considered to be diagenetic.

C^{14} determinations on other Quaternary concretions should throw further light on the absolute rate of concretion-forming processes.

INTRODUCTION AND DESCRIPTION

In June 1956, a fishing trawler recovered a highly fossiliferous concretion from 70 fathoms at a locality (41° 50′S, 174° 26′E) about ten miles southeast of Cape Campbell, New Zealand (fig. 1). In its original form, the concretion (fig. 2) measured roughly 38 cm×35cm×25 cm.[2] The rock consists of a tough, argillo-calcareous matrix containing numerous pelecypod and gastropod shells. The surface of the rock has been heavily attacked by chemical weathering and by Recent boring organisms and is thus very rough and porous. Many of the pits in the surface contained pockets of mud. The freshly-broken matrix is grey in colour, with a brown zone extending 3–4 mm below the weathered surfaces.

In thin section the matrix is seen to be composed mainly of small randomly-oriented prisms of carbonate (0.02–0.05 mm in length), evenly interspersed with cloudy sub-microscopic grains of argillaceous material. Chemical tests show that most of the carbonate is calcite, and that carbonate forms about 50 percent by weight of the matrix. Subangular grains of quartz, corresponding in size to fine sand or silt, are scattered through the matrix. Calcareous shell material is also abundant in thin section; much of it consists of broken shell, apparently derived from valves of the larger species in the rock, but complete tests of Foraminifera and small Gastropoda are quite common.

Two pelecypod species dominate the macrofauna, *Chlamys delicatula* and *Aulacomya maoriana*. The valves nearly always occur singly, but although many are chipped and broken, their ornamentation is well preserved. Most of the *Aulacomya* valves are bleached, and most of the *Chlamys* valves are either bleached or blackened with iron oxide, but traces of original colour are found; a few of the *Aulacomya* valves are pale violet, and occasional *Chlamys* valves show pale pink and yellow bands. The *Chlamys* valves consist very largely of calcite, although a central layer of fibrous aragonite is occasionally present. The *Aulacomya* valves consist dominantly of aragonite recrystallised to calcite in places.

Determinations of the age of both fossils and matrix have been made by the Division of Nuclear Sciences (New Zealand Department of Scientific and Industrial Research), using the C^{14} method. The results are as follows:

Specimen 1 — 30 g of *C. delicatula* and *A. maoriana* valves, extracted from the matrix. Age: 19,500 years ±1000.

[1] N. Z. Oceanographic Institute Contribution No. 38. Manuscript received January 31, 1958.

[2] It was accompanied by a smaller specimen (about 15 cm in diameter) of the same rock-type, which may have been a fragment broken off the larger one. The whole of the smaller specimen, and much of the larger one, have been broken up for fossil extraction.

Specimen 2 70 g of matrix, free as far as possible from enclosed fossils. Age: 27,500 years ± 3000.

ORIGIN AND SIGNIFICANCE OF THE CONCRETION

The matrix of the rock contains only a minor proportion of identifiable organic remains and consists mainly of small prisms of calcite, which appear to have crystallised in place, together with non-calcareous detrital particles. The relative ages of the calcium carbonate in the fossils and in the matrix show that the carbonate in the matrix cannot have been precipitated directly from the sea. It must represent derived organic or detrital carbonate which has undergone recrystallisation.

The rock is clearly derived from a sediment containing argillaceous, calcareous, and silty components. Recrystallisation of carbonate in such an environment usually involves a considerable amount of segregation, carbonate being drawn from the bulk of the sediment and concentrated into nodules which normally range up to a foot or two in diameter and are a great deal harder and tougher than the surrounding material. The specimen under discussion is considered to be a concretionary nodule of this type, formed in a marl or calcareous mudstone containing a small proportion of silt.

The specimen was probably lying in a fairly isolated position on the sea-bed, since it was the only one (except for the smaller fragment previously mentioned) recovered by the trawler during that particular voyage. The irregular surface of the rock proves that it has undergone extensive erosion, in

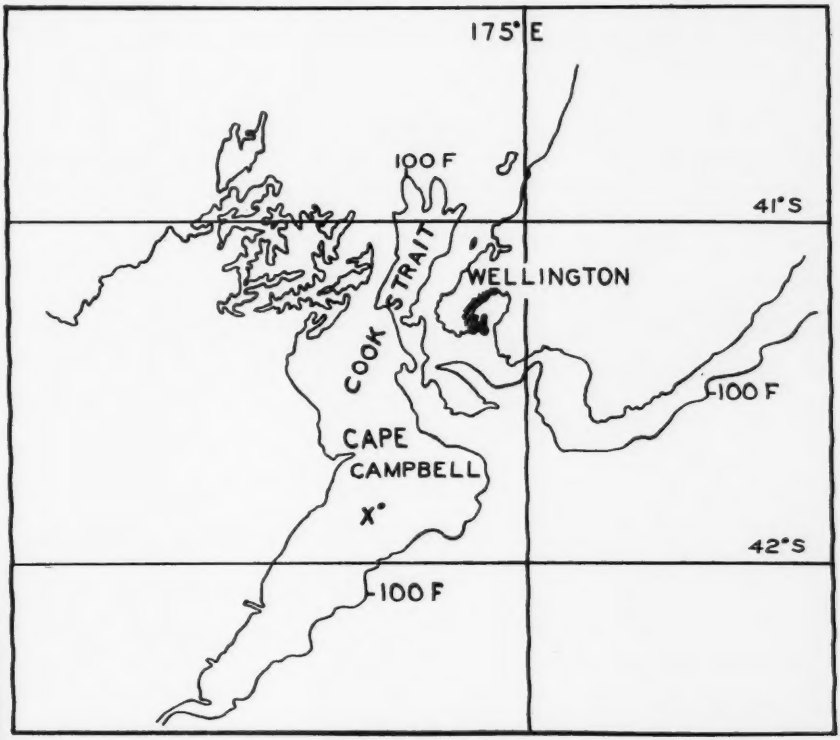

FIG. 1.—Locality map. The 100-fathom isobath corresponds approximately to the edge of the shelf. X marks the position where the concretion was obtained.

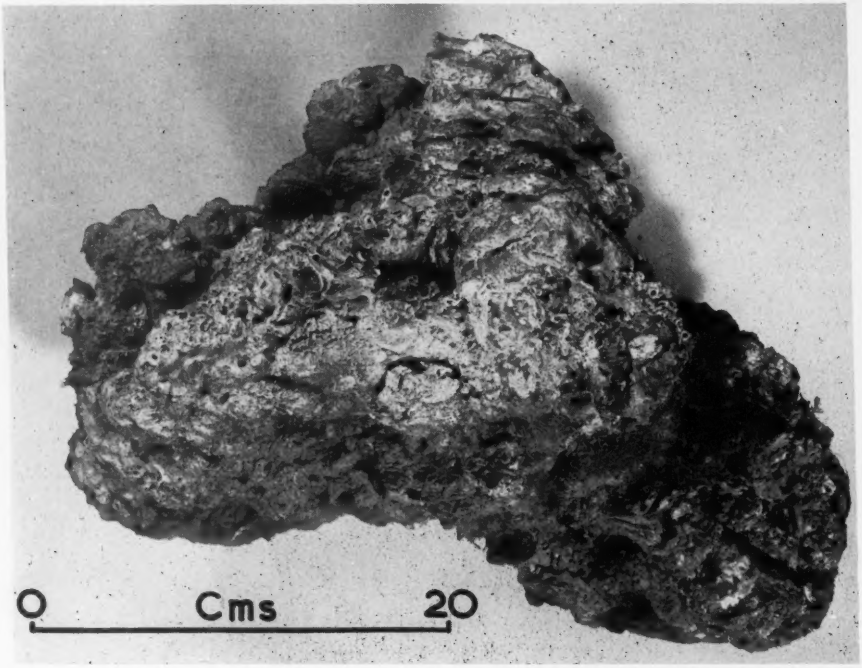

FIG. 2.—The concretion. Photograph by S. N. Beatus.

which chemical and biological attack has played a major part. This line of evidence indicates that it is a relict mass eroded out of some less resistant material such as poorly-compacted silty marl. The presence of mud in the surface pits indicates that this phase of erosion was followed by a phase of sedimentation which may have continued up to the present time.

The molluscs found in the concretion lived about 19,500 years ago. The sediment in which the fauna was originally enclosed (and in which the concretion developed) was probably deposited at the same time or at a slightly later date. The silty and argillaceous components of this sediment must have been derived from pre-existing rocks, and the same applies to the carbonate now forming the matrix of the concretion.

Richardson (1921) listed three types of concretion which might form in sedimentary rocks: (1) contemporaneous concretions formed at the time of deposition of the enclosing sediment, (2) penecontemporaneous concretions formed in the enclosing sediment while it was still soft and unconsolidated, and (3) subsequent concretions formed after consolidation of the enclosing sediment. As it is not possible to determine directly the relationship of the Cape Campbell concretion to the sediment in which it was formed, the specimen can only be assigned to one type or another on the basis of comparison with concretions from other localities whose relationships have been investigated in place.

Most investigators (for example, Tarr, 1921; Weeks, 1957; and Clifton, 1957) have not used the same terms as Richardson in referring to the different types of concretion, the less clumsy "syngenetic," "diagenetic," and "epigenetic" being preferred. Hitherto, these terms have been rather loosely defined, although their usage corresponds roughly (in the order mentioned) to Richardson's three categories. Sometimes only two categories are recognised, Richardson's type (2) being classed with

type (1) or with type (3); this is presumably due to a tacit but unwarranted assumption that conditions within unconsolidated sediment would approximate either to those at the surface (types 1 and 2 being placed together) or to those in consolidated sediment (types 2 and 3 being placed together). The present writer proposes that Richardson's three basic categories be retained, but that the terms "contemporaneous," "penecontemporaneous," and "subsequent" be replaced by "syngenetic," "diagenetic" and "epigenetic," in that order.

Calcareous concretions are commonly found in shales, mudstones, and clays and occur in formations of widely different ages. For instance, they have been described by Tarr (1921) from Pennsylvanian and Cretaceous shales in various parts of North America, by Weeks (1957) from Cretaceous shales in Colombia and from Recent clays in Canada and Greenland, and by Clifton (1957) from the Devonian Ohio shale. These three workers all decided against an epigenetic origin. There is abundant evidence that the concretions formed while the associated mud was still soft and capable of flow. This conclusion is upheld by the frequent presence in the concretions of undeformed organic remains which have been protected from deformation during the compaction of the surrounding sediment.

Tarr (1921) and Weeks (1957) both refer to the concretions they describe as "syngenetic," but it is doubtful whether these examples are really syngenetic in the sense used by the present writer. Any truly syngenetic concretion must be one that formed initially on the surface of the associated sediment, being covered only when its growth was virtually complete. It might be expected that in concretions lying on the sea-bed the form and manner of growth of the free upper surface would be different from that of the lower surface which is resting in contact with the sediment; buried syngenetic concretions, therefore, would probably show an obvious difference between top and bottom. Such an asymmetry is well developed in the syngenetic algal concretions which occur in some limestones (Pettijohn, 1949, Plate 22 top), but is conspicuously absent from the concretions illustrated by Tarr and Weeks. Most of the latter have in effect a plane of symmetry parallel to the bedding.

Ooliths and pisoliths are highly symmetrical, although they are in fact small syngenetic concretions, but these owe their symmetry to the rolling action of currents. The massive concretions described by Tarr and by Weeks can scarcely have been rolled around or even overturned by the weak currents which were presumably responsible for depositing the associated lutites. The general top-and-bottom symmetry indicates that both upper and lower parts of these concretions grew under rather similar conditions, and that they were enclosed by sediment during growth. They would thus be "diagenetic" in the terminology of the present writer. The same argument applies to the concretions described by Clifton (1957), who came to a similar conclusion on other grounds (see below).

This conclusion is upheld by other considerations. Any hard object lying on a mud bottom is usually covered by organic growth within a short space of time, the growth consisting of encrusting organisms such as algae, bryozoa, or molluscs. The concretions under discussion (including those described by Clifton) do not appear to carry encrustations of this type, nor do they show evidence of contemporaneous attack by boring organisms. They also lack the banded or layered structure and the regular orientation of crystals within the layers that characterise the majority of demonstrably syngenetic growths (ooliths, pisoliths, and calc-sinter).

Thus the petrological structure of the concretions described by Tarr, Weeks, and Clifton, the external form of these concretions, their lack of encrusting organisms, and their lack of contemporaneous borings all indicate that the objects did not form at the surface of the associated sediment, i.e. they are not syngenetic. There is equally good reason to believe that the concretions are not epigenetic as previously discussed; they must therefore be diagenetic.

This is the conclusion that Clifton (1957, p. 120) reached with respect to the concretions in the Ohio shale,[3] but his argument against a syngenetic origin appears to be based mainly on the assumption that shale-forming mud would be too soft to

[3] He uses the term "penecontemporaneous" instead of "diagenetic."

bear the weight of a concretion on its surface. This may well be so, although a concretion lying on the surface would tend to compact the mud beneath it, thus gaining additional support. However, the arguments cited above in favour of a diagenetic origin for the concretions uphold Clifton's view.

It seems, then, that calcareous concretions in fine-grained sediments are normally diagenetic, and it is therefore probable that the Cape Campbell concretion is diagenetic. The original external form of the latter cannot, of course, be ascertained, but the specimen is basically similar in petrographic structure to the diagenetic concretions already discussed. The relative ages of the matrix and fauna show, in any case, that the concretion cannot be syngenetic.

Weeks (1957, p. 100–102) presents good evidence for his view that many concretions in argillaceous deposits (including those from Colombia) owe their formation to the anaerobic decomposition of nitrogenous organic matter; this releases ammonia and amines into the environment, which raise the pH and enable carbonate to precipitate from normally undersaturated sea-water permeating the sediments. If the Cape Campbell concretion had been formed by such a process, the carbonate in the matrix should have been contemporaneous with or younger than the shelly fauna. However, the matrix carbonate is significantly older than the fauna, and so the process suggested by Weeks cannot apply in this case. The concretion was apparently formed by the recrystallisation of carbonate particles previously disseminated through the enclosing sediment, these particles having been derived in the first place from older sediments.

It is not yet possible to determine the relative importance of the precipitation of new carbonate and the recrystallisation of old carbonate in the formation of concretions in argillaceous sediments. Neither is it possible to state whether biochemical processes, as a cause of the development of concretions, are more or less important than the purely physical accretion effect (the tendency of large crystals or aggregates to grow at the expense of small ones, which have a higher surface energy). Clifton (1957, p. 120–124) gives a detailed interpretation of the spatial aspects of concretionary growth, but does not suggest an ultimate source for the lime, stating only that the water permeating the sediment was "strongly charged with minerals. . . ." It may be said, however, that the strong tendency of concretions to lie along particular stratigraphical horizons, together with the generally random orientation of calcite crystals within the concretions, can both be simply explained in terms of the accretion principle. Nearly all muds contain some proportion of calcilutite, whether of organic or inorganic origin. As in other types of sediment, there must often be a depositional rhythm due to variations in the environment, resulting in corresponding variations in the calcilutite content of the mud. The zones with a high lime content are the zones in which concretions would be expected to grow, and the calcilutite particles would form readily available nuclei for the precipitation of carbonate wherever a concretion was in the process of growth. The orientation of the calcilutite particles (probably random in most cases) would be inherited by the concretion. The actual position of a concretion within a given stratum probably depends upon an accident such as the presence of a decomposing fish or mollusc; this, for instance, might cause a local rise in pH which could initiate the growth of a concretion (Weeks, 1957, p. 101).

RATE OF FORMATION

As far as the writer is aware, no data have hitherto been published regarding the absolute time needed for the development of calcareous concretions in marine sediments. In the case of the Cape Campbell concretion, however, the C^{14} age of the shelly fauna shows that the concretion formed within the last 19,500 years. At the present time, moreover, sediment is probably being deposited in the area of the shelf where the concretion was obtained. The phase of erosion which exposed the concretion thus did not occur under conditions resembling those of the present day, and can be tentatively assigned to a period of exceptionally low sea-level, such as one of the maxima of the last Glacial Stage. If this is correct, the formation of the concretion was completed before the onset of the

last full-scale glacial maximum (Mankato), which occurred about 12,000 years ago. The evidence thus indicates that the concretion was formed within 7500 years, between 19,500 and 12,000 years ago.

There is widespread stratigraphic evidence that calcareous concretions are able to form in muddy sediments within a few thousand years of deposition. Concretions in clays that are considered to be Flandrian in age have been described by Bartrum (1917) from the muddy infilling of Auckland Harbour, by Etheridge and McCulloch (1917) from coastal clays in Queensland, and by Weeks (1957) from coastal clays in Greenland and Canada. These Quaternary concretions are probably diagenetic. They are more irregular in shape than the older concretions described in a previous section, but this might simply mean that they are in an earlier stage of growth. They are basically similar to the older concretions from a petrographic point of view, the calcite with them forming small, randomly-oriented prisms. C^{14} determinations on some of these examples would yield interesting results, and the writer has submitted specimens from Auckland Harbour for study by the N. Z. Division of Nuclear Sciences.

Finlay and Benson (1950, footnote to p. 270) mention the recovery of calcified crockery from the mud near the wharves at Woolloomooloo Harbour, Sydney. The carbonate must, in this case, have been precipitated within a few decades. The material, however, is not figured or described in detail in the above paper, and it is thus uncertain whether the carbonate is syngenetic or diagenetic.

CONCLUSIONS

Concretions in marine sediments may be divided into three types:

1.—Syngenetic, formed on the sea-bed at the time of deposition of the enclosing sediment.

2.—Diagenetic, formed (probably within a few thousand years) after the deposition of the enclosing sediment, but before the various processes of compaction and consolidation were complete.

3.—Epigenetic, formed after the compaction and consolidation of the enclosing sediment.

On this classification, the Cape Campbell concretion is considered to be diagenetic. It was formed within the last 19,500 years, and its formation probably occupied 7500 years or less, the carbonate in the matrix being derived not directly from the seawater but from organic or detrital carbonate distributed through the enclosing sediments.

ACKNOWLEDGEMENTS

The writer wishes to thank Mr. T. A. Rafter and Mr. G. J. Fergusson for the C^{14} age determinations quoted in this paper.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 372–375
FIGS. 1–3, SEPTEMBER, 1958

THE ZIG-ZAG NATURE OF TYPE I AND TYPE IV CURVES[1]

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ABSTRACT

Many distributions of sedimentologic (and other geologic) data are non-Gaussian. Of nine examples studied in detail, seven are Pearson's Type IV, and two are Pearson's Type I. Each plots as a zig-zag line composed of two or more straight-line segments on probability paper. The zig-zag pattern appears to be caused by a deficiency of particles of certain sizes. This is perhaps a general, rather than a local, condition, as has been pointed out by many workers.

Probability paper has proven very useful in various geological studies, particularly sedimentology. The Gaussian (that is, so-called "normal") distribution plots as a straight line on probability paper. For many samples this permits an easy, quick evaluation of curves which otherwise would not have nearly so much obvious meaning.

A number of years ago it began to be plain that many curves will not plot as straight lines on probability paper (Bagnold, 1941, p. 118; see also Inman, 1952, for a use of Pearsonian curves, and Otto, 1939, for Gram-Charlier series). These departures from the Gaussian curve are of considerable interest. They commonly consist of several connected straight line segments. One might reason that each segment represents a Gaussian distribution, and that the entire sample is a mixture of two or more otherwise "normal" samples. This basic idea is not examined in the present paper.

It has been known for many years by statisticians (Kendall, 1947, p. 132), that there is nothing particularly normal about the "normal" curve and nothing particularly sinister about any other curve type. The Gaussian distribution owes its present pre-eminence primarily to the fact that, as a degenerate form of something else, it is more easily manipulated. In the Pearson system (numbered according to Kendall), Types I, IV and VI are the important curves. The Gaussian curve, which does not fit into any of these three types, is less typical although more easily handled theoretically. The three fundamental types, plus the Gaussian, may be distinguished as follows:

1. Calculate the first four moments about the mean (U_1, U_2, U_3, U_4).
2. Obtain the parameters, $\beta_1 = U_3^2/U_2^3$ and $\beta_2 = U_4/U_2^2$.
3. Solve the three equations,
$$\left.\begin{aligned} b_0 &= U_2(4\beta_2 - 3\beta_1)/B \\ b_1 &= \sqrt{U_2}\sqrt{\beta_1(\beta_2+3)}/B \\ b_2 &= (2\beta_2 - 3\beta_1 - 6)/B \end{aligned}\right\} \text{ where } B = 10\beta_2 - 12\beta_1 - 18.$$
4. Take the roots of $b_2x^2 + b_1x + b_0$.
5. Calculate Pearson's criterion,
$$K = \frac{\beta_1(\beta_2+3)^2}{4(2\beta_2 - 3\beta_1 - 6)(4\beta_2 - 3\beta_1)}.$$

The following tabulation permits the use of the roots (step 4) and the criterion (step 5) in identifying certain curves (see Fig. 3).

Roots	*Criterion*	*Type*
real; opposite sign	negative	I
imaginary	between zero and one	IV
real; same sign	one or greater	VI
$b_1 = b_2 = 0$		Gaussian

The curves shown in figure 1 were chosen as representative from a much larger group. Each curve in the group has at least two connected straight line segments. Some of them have five or more. The intersections of the straight segments fall in various parts of the chart. A curve having only one intersection, the intersection being close to the 99% line, may appear to belong to the Gaussian category. In many instances the zig-zag nature of the line would not have been obvious had the curve not been plotted on probability paper. The range of sizes present in many sediments is not sufficiently great for the student to be able to determine whether the curve is Gaussian or one of Pearson's types.

[1] Manuscript received March 10, 1958.

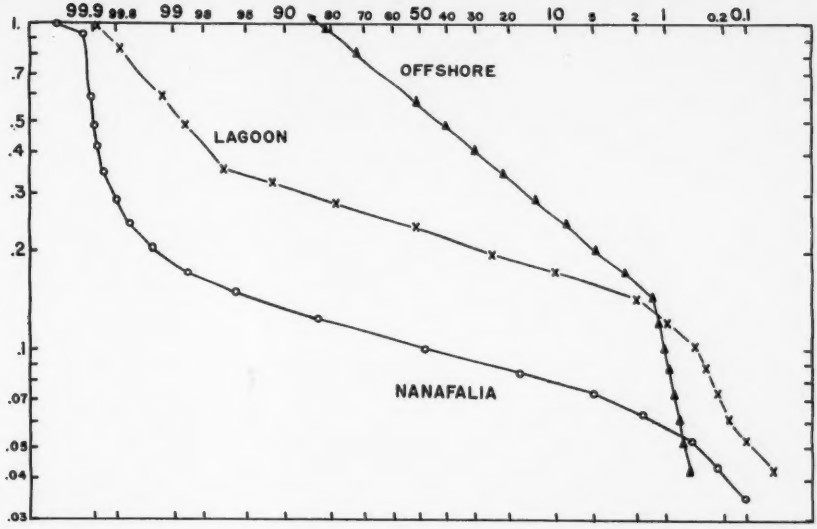

FIG. 1.—Three examples of the Type IV curve. The lagoon and offshore samples were collected along the Florida coast, southwest of Tallahassee. The Nanafalia sample came from the outcrop of that formation near Andersonville, Georgia. These three samples are varied but representative examples chosen from many others. Note that the Nanafalia line is gently curved in the neighborhood of 99%, rather than showing a sharp inflection as the lagoon sample does near 98%. This not-so-typical portion of the curve may be caused by the fact that the coarser part of the sample is made up largely of clay pellets.

The identification of Type I and Type IV curves as such does not "explain away" the zig-zag pattern. The intersections are real, and the changes in line slope must have a meaning of some kind. Such an explanation is already available in the literature.

The most common intersection on the curves studied to date is C (fig. 2). The next most common is D. The straight line segment, CD, is plotted against the variable scale of probability paper and therefore must be read with caution. Near C it represents a *decrease* from the values which would be obtained by projecting BC as a straight line. Near D it represents an *increase* in values. It is suggestive, therefore, of the minimum between two modes. (In actual practice, however, class intervals may be too wide for the secondary mode to be obvious.)

Likewise, the line AB suggests a minimum, with the part of the line closest to B indicating a *decrease* relative to the projection of BC, and the part closest to A indicating an *increase*.

For the sediments studied (approximately 400 different samples), the following average values were obtained for the four intersections:

A. (Not clear; few data)
B. 0.75 mm (3/4 mm)
C. 0.11 mm (1/9 mm)
D. 0.05 mm (1/20 mm)

The two minima therefore appear to have values of, approximately:

AB. 1 to 2 mm
CD. 0.06 to 0.08 mm (About 1/15 mm)

Groot (1955) observed that the intersection here labelled C might be explained in terms of the division between traction (that is, bed) load and suspension (that is, wash) load. He presented more than a dozen curves showing C-points. From the sample plots available for the present study it is concluded that the minimum line suggested by CD is more nearly the division between wash and bed load materials. The AB

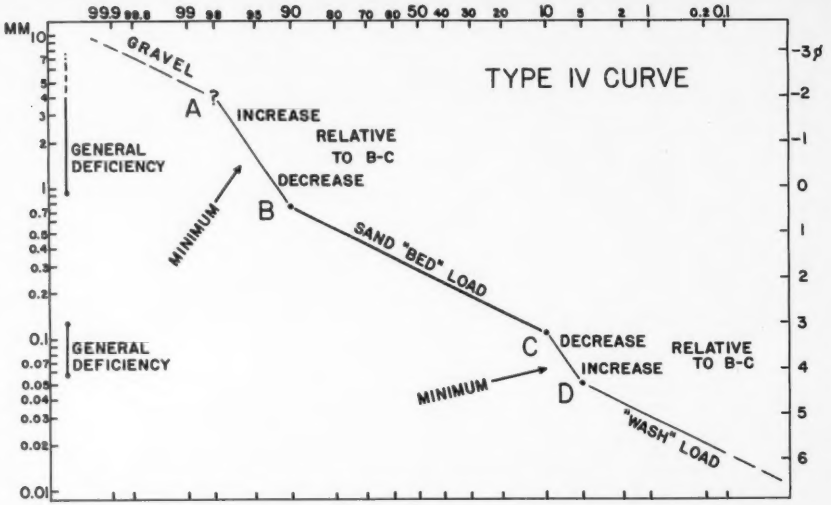

FIG. 2.—Diagrammatic example of the Type IV curve. The three modes may represent gravel bed load, sand bed load, and wash load. The inflection points, A, B, C, and D, are discussed in the text. The minima (between modes) occur along the segments AB and CD; these minima correspond approximately with the observed deficiencies of sediment in the sizes indicated. The locations chosen for B, C, and D are average values obtained from 48 Type IV curves selected from approximately 400 analyses; the data for A were not clear, partly because few gravels were included. Determination of inflection points for many other samples might alter the values used here. The words "increase" and "decrease" refer to a comparison of the plotted line with the extension of the BC segment; that is, the 1-mm fraction actually contains less material ("decrease") than would be indicated by an extension of BC toward the upper left.

minimum appears to separate the gravel bed materials from the sand bed load.

The fact that detrital sediments in general appear to be deficient in certain sizes has been known for many years. Pettijohn (1949, p. 41–45) summarized previous work on this problem. His summary, covering thousands of samples ranging from silt to gravel, made clear the deficiency of materials in the 0.03–0.12 and 1.0–8.0 mm sizes.

In the light of Pettijohn's summary, deficiencies in the same size ranges in specific samples are explained in general rather than local or environmental terms. In other words, weathering and transportation appear to provide three types of material:

1. Blocks (perhaps composed of many grains).
2. Grains (that is, quartz grains).
3. Flakes or platelets (that is, clay).

Blocks appear to be relatively uncommon below about 2 mm. Platelets appear to be uncommon above about 0.06 mm. And single grains seem to fall predominantly in the range, 0.8 to 1 mm. The combination plots as a zig-zag rather than a straight line on probability paper. However, the sediment obtained by the attrition of a limestone or chert gravel might have quite different characteristics.

A specific sediment which shows the C-point includes the BC segment (sand bed load) in whole or part and at least part of the CD transition to wash load. If the D-point also appears, then an appreciable fraction of the wash load has been trapped along with the sand.

The absence of the B-point probably means that no gravels were available for transportation to, and deposition at, the sample site.

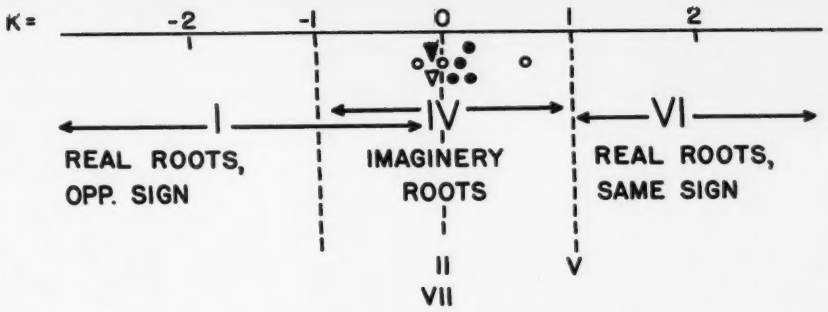

FIG. 3.—Chart showing Pearson's criterion, K, and the nature of the roots (see Step 4 in the text) for nine samples. The important curves, Type I, Type IV, and Type VI, have values of K indicated by the arrows; Type I and Type IV overlap, but can be distinguished by the differences in roots. At least three of Pearson's less general cases (Type II, Type V, and Type VII) fall on precise values of K, as indicated at the bottom of the chart. Two of the samples (triangles) are Type I curves; they are obviously bimodal. Seven of the samples (circles) are Type IV curves. The solid symbols represent sediments, the open symbols geomorphic data. The sediments include the three given in figure 1. The Gaussian distribution is not conveniently represented on this chart.

It is, therefore, thought that no specific explanation is needed for the zig-zag pattern shown by any single Type I or Type IV curve. Rather, special attention should be given to any plot which appears to be perfectly Gaussian to explain why the various inflection points are missing.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 376–378
SEPTEMBER, 1958

NOTES

SOME APPARENTLY MISUNDERSTOOD ASPECTS CONCERNING STYLOLITES[1]

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It appears that stylolites are still among the least understood of the common structures that appear prominently in the calcareous sedimentary rocks. In one of our recent textbooks, "Sedimentary Rocks" (Pettijohn, 1949 p. 156), we find these statements in his description of stylolites—"the teeth-like projections of one side fit into sockets of like dimensions on the other. In cross section the stylolitic surface resembles somewhat the tracings of a stylus, hence the term *stylolite* is applied to this feature." In the second edition of the same work (Pettijohn, 1957, p. 213) we find that the description of stylolites has been changed somewhat to read as follows: "the teeth-like projections of one side fit into sockets of like dimensions on the other. In cross section the stylolitic surface resembles a suture (Ger.: Drucksuturen) or the tracing of a stylus—an oscillogram."

There is but slight difference in these two statements, although in the latter the statement concerning the origin of the term stylolite is omitted although possibly inferred. In the first edition the origin of the term stylolite is assumed to be from the Latin *Stilus*, an instrument for writing, rather than from the Greek *Stylos* a pillar or column. The term stylolite was given to the structure by Klöden (1828, p. 58) who believed the structure to be the fossilized remains of a distinct organic species to which he gave the name "*stylolithes sulcatus*" meaning a fluted or furrowed column. The name so aptly describes the physical appearance of the columnar type of the structure that it has been universally accepted by geologist. The word *sulcatus*, however, comes from the Latin and means furrowed.

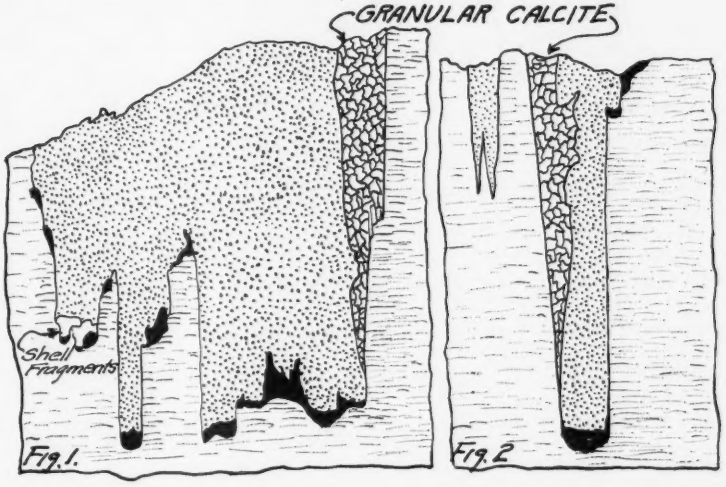
Concerning the first part of the statement in which Pettijohn stated that the "teeth-like projections on one side fit into sockets of *like dimensions* (italics mine, B.M.S.) on the other" appears to be a necessary situation of utmost importance to those committed to the pressure-solution theory for the origin of stylolites by solution in an already hardened rock. If one searches diligently, he will find numerous instances where the columns are smaller than the sockets. The presence of cavities along stylolite seams has been described by Shaub (1939, p. 56. fig. 4) who has also reported open spaces between stylolite columns and their walls (Shaub, 1953, p. 261, fig. 1). In these instances the two sides of the seam or the column and its socket were originally in contact and were pulled apart by contraction within the respective materials accompanying the loss of water during consolidation of the lime muds. The vertical component during compaction is usually completely compensated by the weight of the overlying sediments while the horizontal stresses, developed by the loss of water, have the ability in some instances to open up cavities along planes of discontinuity, as between the opposite sides of stylolite seams when in a vertical position and between the sides of the columns and their walls. In the cases cited above the customary furrows are on both the column and the wall or on the two sides of the seams where they were developed originally by the relative movement when these parts were in contact and still plastic.

The cavities thus formed may remain open with the striae showing on the two sides or the striae may be partly or entirely covered with dolomite or calcite depending upon the composition of the rock in which the stylolite seams or columns occur. In some instances the cavities or open spaces along the column and socket are completely filled with rather coarse granular calcite as shown in figures 1 and 2.

[1] Manuscript received March 26, 1958.

The areas of coarse calcite in these illustrations represent sections across openings which were originally formed between the columns and the analogous sockets. In such instances the columns and sockets are of *unequal dimensions*, the column having shrunk away from the socket. There is at present no evidence that the walls had been displaced after the rock had consolidated as there is no fracturing or other disturbance that can be associated with post consolidation displacement.

It would be difficult to see how Pettijohn could have overlooked the description of the cavities along stylolite seams and about the stylolite columns for he has listed the titles of the papers in which this phenomenon was described in the bibliography at the end of the chapter concerned with stylolites. On page 215 (second edition) he states that "though a solution-pressure origin is a well-established conclusion, the *mechanism* by which solution, acting under directed pressure, makes a stylolite is far from clear." A serious student might ponder over the statements that a theory can be "well established" and yet the mechanism by which it operates is "far from clear." In a footnote (page 215) Pettijohn states that "there appear to be, as always, some who dissent" without giving any statements about the reasons for dissention. It would appear that the presentation of the stylolite problem in the above mentioned text is rather unfair to the beginning student whose research time is limited and who has to accept from a recent text on sedimentary phenomena such biased and incorrect statements of the facts as presented there concerning one of the oldest and most conspicuous of the phenomena associated with highly calcareous sediments.



FIGS. 1 and 2.—These illustrations were carefully traced from photographs in order to be assured that the reproductions in printing will clearly delineate the areas of shrinkage between the columns and the walls. The spaces or voids developed are now filled with granular calcite. Specimens from the Bedford, Indiana, limestone area. The stippled areas represent the sections of stylolite columns; black shows the clay along the seams and columns; patched areas represent granular calcite which fills voids between the columns and their sockets.

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COMMENTS ON "SOME APPARENTLY MISUNDERSTOOD ASPECTS *CONCERNING STYLOLITES*"

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That the term "stylolite" was derived from the Greek *stylos*, a pillar or column, has been noted by Stockdale (1922, p. 10) and others (Twenhofel, 1932, p. 733). Though fully conversant with these references, the writer inadvertently attributed the term to the resemblance between the appearance of a stylolite in polished cross-sections to the tracing of a stylus (1949, p. 156). The resemblance is undeniable, but the origin of the term is not related to this fact and hence the erroneous statement cited was omitted in later publication (1957, p. 213). There is no need to flog a dead horse.

Prof. Shaub is perturbed that "a theory can be 'well established' and yet the mechanism by which it operates is 'far from clear'." One can readily establish by medical evidence that a particular person met his death by violence but be unable to say just how and why this was accomplished. And geologists generally agree that the glacial origin of till, drumlins, and related features is considered a well-established theory although there is much uncertainty as to how or in what manner glaciers deposit till or construct drumlins. And so too the pressure-solution origin of stylolites has now been demonstrated beyond any reasonable doubt although the exact mechanics of the action remains unclear.

The statement that the granular calcite, shown in Prof. Shaub's sketches, "fills voids between the columns and their sockets" and constitutes proof of "shrinkage between the column and the walls" is a subjective judgement and nothing more. Granular calcite *may* be a cavity filling but as every student of carbonate rocks knows, it may also form by replacement and by recrystallization.

Whether the statement on stylolites published in *Sedimentary Rocks* "is rather unfair to the beginning student . . . who has to accept . . . such biased and incorrect statements of the facts as presented there . . . ", the members of the geological profession will have to decide for themselves.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 379–380
SEPTEMBER, 1958

REVIEW

The Admirable Discourses of Bernard Palissy, translated by Aurèle La Rocque. Pp. vi+264; 6¼×9¼ in., cloth. With Foreword and Introduction by the translator. University of Illinois Press, Urbana, Ill., 1957. Price $5.50.

Under the stimulus of Dr. George White's enthusiasm for the history of geology, the University of Illinois Press has recently published a worthy companion volume to its reprint of Playfair's *Illustrations of the Huttonian Theory of the Earth* (Jour Sedimentary Petrology, v. 27, p. 351–352). If we find in the writings of Hutton and Playfair some roots of modern geomorphology and sedimentology, others may be discovered much deeper in history. Among the most remarkable of earlier writers on these subjects was Bernard Palissy, who lived from about 1510 to 1590 and who made the first enamelled pottery (faience) in France. Widely travelled within his native land, trained in certain aspects of civil engineering, and habitually using shellfish and other animals as models in decorating his pottery, Palissy wrote two books, the second and more important of which has now been admirably translated and annotated by Professor La Rocque.

Palissy assumed as his basic premise that God had supplied all the materials now found on earth at the time of the Creation, though their form was constantly changing. The vast quantities of rocks destroyed by waves, carried by winds, dissolved by rainwater, and shattered by frost were therefore presumed to be replaced by the growth of rocks elsewhere. Such growth was controlled by the availability of "congelative water," in which different types of "salts" were dissolved. The type of "salt" controlled amongst other things, the colour, the transparency, the hardness, the cohesion, the porosity, the crystalline structure, and the the weight of the growing stone mass. Palissy insisted that cold, not heat, was required for the growth of the hardest rocks, and that the only agent capable of breaking down congealed rocks was ordinary (as opposed to "congelative") water.

Palissy applied this line of reasoning to the interbedding of well congealed and poorly congealed rocks which he had observed in the Paris and Aquitaine Basins. He explained that rainwater, percolating downwards through soil and regolith, dissolved large quantities of "salts." When its downward progress was stopped by solid rock, the water would congeal to rock itself, thus forming a new stratum. If the overburden were "muddy," proper congelation would be prevented and a marked break between the old and the new strata would occur. Vertical fissures also represented local differences in congelative conditions. Congelation could also take place in crevices within the bedrock, producing deposits of iron, silver, lead, tin, or gold, depending on the type of "salt" involved.

Palissy believed that sand grains often had a purer "salt" content than their "muddy" cement and consequently tended to stand out in relief on exposed faces of sandstone. Cohesive clay was often found at the *base* of a series of strata because its impermeability made for good congelative conditions above.

Where the upper surface of an impervious layer, such as bedrock itself, was tilted, congelation would not occur at all. The water would, rather, flow along such a surface until it emerged into the open air as a spring. Palissy was insistent that the presence of bedrock was essential to the occurrence of most springs and that spring water itself was merely percolated rainwater. For artesian conditions to prevail, the source of the water must have been higher than the base of the well or higher than the point where the water emerged spontaneously from sandy ground.

Many other topics of interest to sedimentologists and geomorphologists were discussed by Palissy in his *Admirable Discourses*. Of these, the following deserve special mention: the preservation and organic origin of fossils, former inland transgressions of the sea (up to 100 miles), chemical weathering, the formation of river ice, waves and shore erosion, the hydrologic cycle, the checking of fluvial erosion by reforestation, earthquakes and block-faulting, and specifications for constructing different types of reservoirs and a soil auger. Palissy

knew that rivers deepened their channels and could wash away any hills made of unconsolidated material, but he failed to grasp that river erosion could carve whole valleys and landscapes out of solid rock.

Through the *Admirable Discourses* Palissy emphasizes the virtues of careful observation, logical reasoning, laboratory experiments, field work, and the collection of specimens. Indeed, almost the whole book consists of a dialogue between Theory and Practice, the latter representing Palissy himself and the former his own conception of traditional scientific thought. The fact that Palissy claimed that Nature was his only teacher is the chief topic of discussion in Professor La Rocque's illuminating Introduction to the new translation.

La Rocque states (p. 15): "By great good fortune rather than planned design, observation almost always comes first with him, deduction and explanation second. In this approach to natural phenomena lies Palissy's greatest claim to originality." No doubt this is true. Palissy's methods of exposition are also vigorous and lucid, leaving his reader with the impression of a powerful independent mind.

On the other hand, it should be borne in mind that Palissy was in constant touch, both in Saintes and in Paris, with many learned and travelled men, including mathematicians, physicians, surgeons, apothecaries, lawyers, merchants, poets, clergymen, courtiers, and a certain Jacques Besson, who wrote a book of his own on the origin of springs. Not only that, but for nine years Palissy presented to such men an annual course of lectures, covering the subject matter of the *Admirable Discourses*, during which he invited questions and criticism. Leaving aside his unconvincing assertion that no one ever contradicted his views, we must surely conclude that he picked up a great mass of information and ideas through such contacts, thus by-passing his claimed ignorance of Latin and Greek. Moreover, the fact that almost all of the works which he mentions in the *Discourses* had already been translated into French adds further to suspicion that Palissy learned more from his contemporaries and predecessors than he or Professor La Rocque are willing to admit.

Although La Rocque presents a detailed and on the whole well-balanced defense of Palissy's originality, the reviewer would have liked to see contemporary authorities quoted for such statements as ". . . such an attitude struck the learned doctors of the University of Paris like a thunderbolt" (p. 14) and "There is no doubt that Palissy's approach to science struck his contemporaries as original" (p. 14). Clearly, the relationships between Palissy and the leading academic figures of his day need further study.

But despite such minor objections La Rocque's work must be recommended in the strongest terms to all who are interested in the history of science. The translation keeps very closely to the original text, even to the puctuation, and has obviously been executed and annotated with patient care. The publishers are to be congratulated on the legibility and general appearance of this valuable work.

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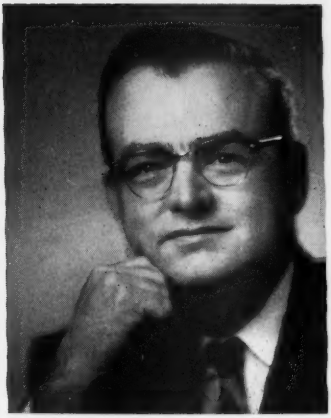
JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 28, NO. 3, PP. 381–383
6 FIGS., SEPTEMBER, 1958

SOCIETY RECORDS AND ACTIVITIES

NOMINEES FOR S.E.P.M. PRESIDENT, 1959–1960



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Born, July 1, 1914, Kansas City, Missouri

Academic Training

Univ. of Kansas City, A.B., '36
Univ. of Missouri, M.A., '38; Ph.D., '40

Experience

1939–43	Missouri School of Mines, Rolla, Mo., instructor
1943–44	Missouri School of Mines, Rolla, Mo., assistant professor
1944–48	Stanolind Oil & Gas Co. (Pan American Petr. Corp.), Midland and Wichita Falls, Tex., geol., sen. geol., and dist. geol.
1948–	Univ. of Texas, Austin, Tex., prof., and chm. of dept.

Publications.—Petroleum geology, micropaleontology, paleoecology, educational techniques, and stratigraphy

Professional Affiliations (National)

Geological Society of America, fellow
American Association for the Advancement of Science, fellow
American Association of Petroleum Geologists, member
Society of Petroleum Engineers, member
Sigma Xi, Sigma Gamma Epsilon

S.E.P.M. Activity (Member, '49)

1952–58	Secretary-Treasurer
1958–59	Chairman, Publications Committee

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Academic Training

State University of Iowa, B.A., '34; M.S., '35; Ph.D., '38

Experience

1939	U. S. Geol. Survey, jr. geol.
1938–40	State University of Iowa, Iowa City, Iowa, res. assoc.
1940–41	Oklahoma A. & M. College (Oklahoma State University), Stillwater, Oklahoma, instructor
1941–44	Shell Oil Co., Tyler, Texas, geol.
1944–46	R. Lacy Inc., Longview, Texas, geol.
1946	Union Producing Co., Tyler, Texas, geol.
1947–49	Creole Petroleum Corp., Maracaibo, Venezuela, dist. geol.
1949–53	Arabian-American Oil Co., Dhahran, Saudi Arabia, div. geol.
1953–58	State University of Iowa, Iowa City, Iowa, assoc. prof. and prof. of geol.

Publications.—Paleontological studies on cephalopods and conodonts

Professional Affiliations (National)

Geological Society of America, fellow
American Association of Petroleum Geologists, member
Association of Geology Teachers, member
Sigma Xi

S.E.P.M. Activity (Associate, '50; Member, '55)

1955–57	Editor,—*Journal of Paleontology*
1957–58	Chairman, Publications Committee
1958–59	Vice-President

NOMINEES FOR S.E.P.M. VICE-PRESIDENT, 1959–1960

DAN E. FERAY

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Academic Training

Univ. of Tulsa, B.S., '39
Univ. of Illinois, M.S., '40
Univ. of Wisconsin, Ph.D., '48

Experience

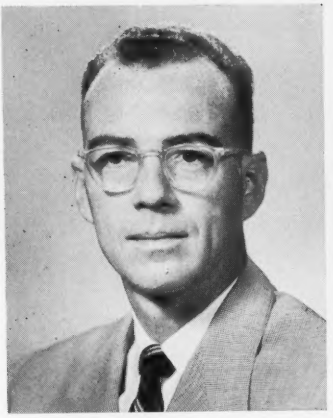
1942–46	U. S. Army Air Force
1946–47	University of Wisconsin, Madison, Wisconsin, lab. instructor
1947–49	Bureau of Economic Geology, University of Texas, Austin, Texas, field asst. and dir. of well sample lib.
1949–51	University of Tulsa, Tulsa, Oklahoma, assoc. prof.
1951–57	Magnolia Field Research Laboratory, Dallas, Texas, in charge of geol. res.
1957–	Feray Geological Service, Dallas, Texas, owner, and Southern Methodist University, Dallas, Texas, Prof. Magnolia Chair Geol.

Publications.—Sedimentation

Professional Affiliations (National)

American Association of Petroleum Geologists, member

S.E.P.M. Activity (Member, '46)



WILLIAM J. PLUMLEY

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Academic Training

Univ. of Chicago, B.S., '40; M.S., '41; Prof. Certificate in Meteorology, '42; Ph.D., '48

Experience

1941	Carter Oil Co., geol. trainee
1942–46	U. S. Army Air Force, meteorology
1946	Carter Oil Co., geol.
1948–	California Research Corporation, La Habra, Calif., res. geol.

Publications.—Sediment transport, carbonate rock petrology, sediment textural analysis

Professional Affiliations (National)

American Association of Petroleum Geologists, member

S.E.P.M. Activity (Member, '55)

1956–58	Research Committee member
1957	Co-chairman of Symposium on Sedimentary Geochemistry

NOMINEES FOR S.E.P.M. SECRETARY-TREASURER, 1959–1960

JOHN IMBRIE

Professor of Geology and Geological Research, Columbia University, New York, N. Y.
Born, July 4, 1925, Penn Yan, New York

Academic Training

Princeton University, B.A., '48
Yale University, M.S., '50; Ph.D., '51

Experience

1951–52	University of Kansas, Lawrence, Kans., research and teaching
1952–57	State Geological Survey, Lawrence, Kansas. (part time)
1952–	Columbia University, New York, N. Y., prof. of geol. and geol. res.

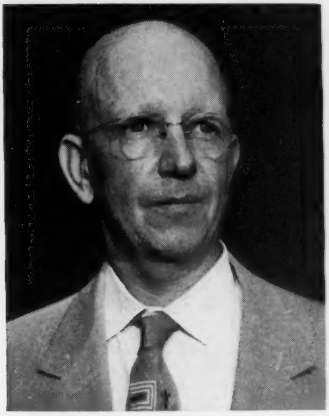
Publications.—Systematic paleontology, biometrics, general paleontology, biostratigraphy, shoal-water geology of Bahamas

Professional Affiliations (*National*), member

Geological Society of America
American Association of Petroleum Geologists, member
Society of Systematic Zoologists
Paleontological Society, member
Sigma Xi

S.E.P.M. Activity (Associate, '52; Member, '54)

Participation in various programs, including the Symposium on Geochemistry of Sediments, and the 1954 Symposium on Fauna and Facies



RAYMOND D. WOODS

Acting Research Coordinator, Geologic Research Section, Humble Oil & Refining Company, Houston, Texas
Born, September 14, 1910, Evangeline, Louisiana

Academic Training

Univ. of Texas, B.A., '31; M.A., '34
Louisiana State University, graduate study '47

Experience

1931–34	University of Texas, Austin, Tex., lab. instr.
1934–42	Humble Oil & Refining Co., Houston, Tex., paleontologist
1936–42	University of Houston, Houston, Tex., geol. instr. (night)
1942–45	U. S. Army Corps of Engineers
1945–47	Humble Oil & Refining Co., Houston, Tex., paleontologist
1947–48	Louisiana State University, Baton Rouge, La., lab. instr.
1948–	Humble Oil & Refining Co., Houston, Tex., paleontologist, stratigrapher, coordinator of research, acting coordinator of research

Publications.—Structure and stratigraphy of Gulf Coast area

Professional Affiliations (*National*)

American Association of Petroleum Geologists, member
Geological Society of America, fellow
American Geophysical Union, member

S.E.P.M. Activity (Member, '46)

Erratum

The following correction should be made in the article by Briggs, Louis I., 1958, Evaporite facies: Jour. Sedimentary Petrology, V. 28, p. 46–56.

The last sentence of paragraph 2, column 1, page 49 should read; "2.—The evaporite pattern, in the simplified hypothetical case presented, is arcuate about an inlet, the *concave* side being toward the inlet."

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